Senior Thesis

An Investigation of the Hydraulic Conductivity of the Scioto River Streambed at the South Well Field, Southern Franklin County, Ohio

by
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INTRODUCTION

Extensive glacial-outwash deposits underlie much of southern Franklin County. These deposits, consisting predominantly of sand and gravel with minor amounts of clay, comprise a major unconfined aquifer. The City of Columbus developed the South Well Field in this aquifer to provide a portion of the City's municipal water supply. A 1976 report to the City of Columbus by Alden E. Stilson and Associates determined that the South Well Field was capable of producing 50 million gallons per day (Mgd) based on recharge via induced stream infiltration from the Scioto River. Three radial collector wells were built along the Scioto River (and a fourth along Big Walnut Creek) to maximize induced stream infiltration.

Radial collector wells work in the following manner. A large diameter caisson is constructed next to a river (in this case the Scioto River) and a series of lateral pipes radiate outward from the central caisson under the streambed to induce the flow of river water into the aquifer and into the wells. This helps recharge the aquifer as water is being withdrawn from it, thus allowing a higher sustained yield of ground water. This type of system functions best in geologic materials of high permeability.

Stilson (1976) estimated that over eighty percent of the pumpage from the wells would originate as induced infiltration from the Scioto River. In reality, this figure is much lower and has been continually revised downward by other investigators (Table 1). A computer model by Eberts (1987) determined that only thirteen percent of the pumpage originated from the Scioto River. This estimate is also corroborated by geochemical evidence by de Roche and Razem (1984), who determined that approximately twenty percent of the pumpage in Collector Well 101 (see Figure 1) is river water.

Although the aquifer may have a high permeability, no study has ever been performed to measure the permeability of the streambed directly. Vertical hydraulic conductivity, a function of permeability and fluid properties, is an important factor in induced stream infiltration. If the streambed is relatively impermeable, the stream is effectively isolated from the aquifer as a source of recharge.

Purpose and Scope

The purposes of this study are to present and interpret vertical hydraulic conductivity data obtained from direct measurements of the Scioto River streambed with a seepage meter and piezometer. These data will be analyzed and compared with data from previous studies.

Location of Study Area

The study area is located in southern Franklin County, Ohio, in portions of Hamilton Township, Jackson Township, and the City of Columbus. The study area includes the Scioto River from Interstate Route 270 to State Route 665 (Figure 1).

Table 1
Percentage of Pumpage Derived from Scioto River

Study	Year	%	Method
Stilson	1976	81	Hydrologic Budget*
Stowe	1979	74	Hydrologic Budget*
Weiss & Razem	1980	70	Computer Model
Razem	1983	32	Computer Model
de Roche & Razem	1984	20	Geochemical Model
Eberts	1987	13	Computer Model

*Based in part on pumping-test data

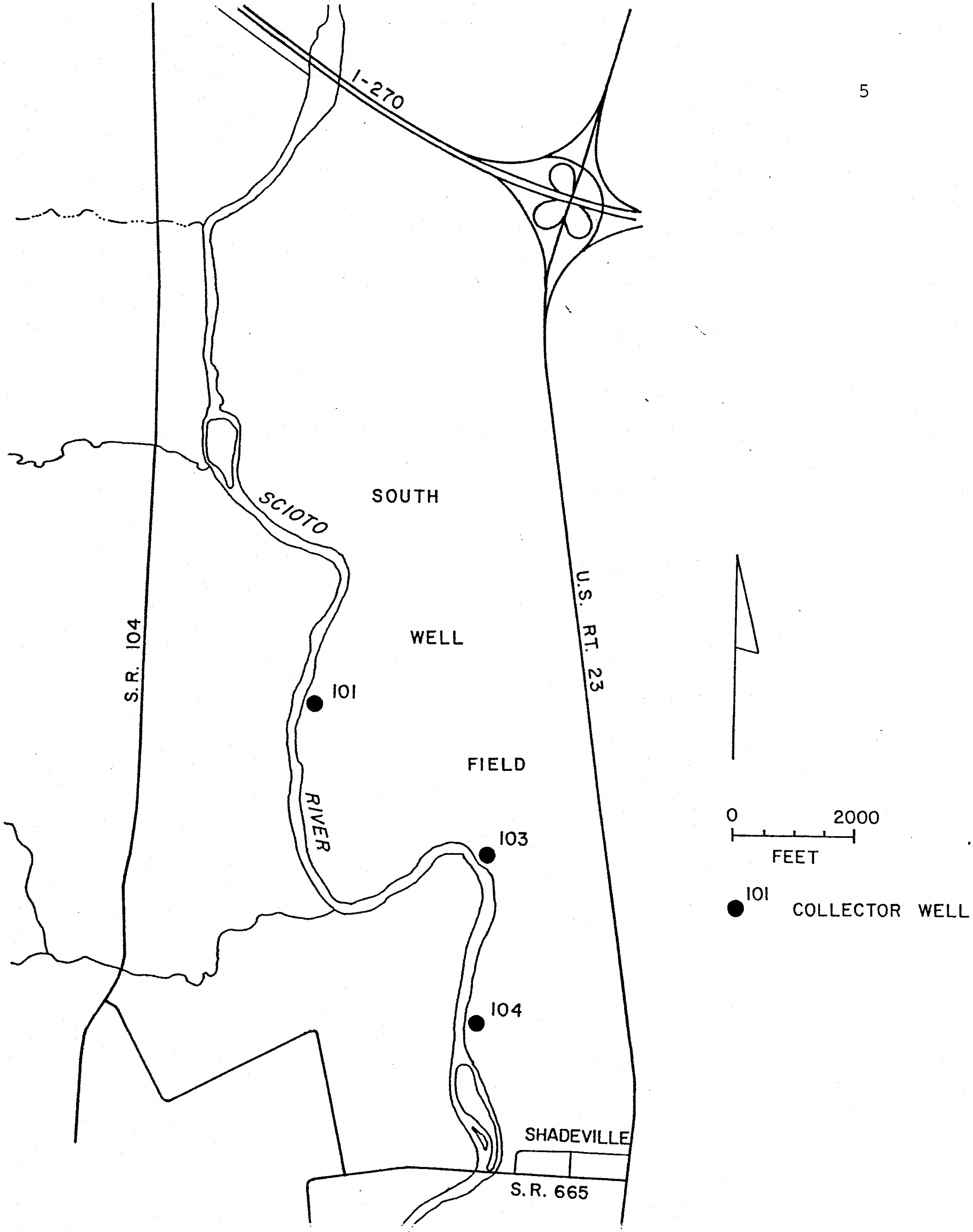


Fig. 1 Map of Study Area

GEOLOGIC SETTING

South-central Franklin County lies within the Central Lowlands physiographic province (Fenneman, 1938). The present topography is developed in Wisconsinan Age glacial deposits, and has been modified by post-glacial drainage (Figure 2). Topography in the vincinity of the study area ranges from mildly sloping to moderately rolling with a maximum relief of approximately 150 feet. The Scioto River valley in the study area varies from about 3,500 feet to 5,500 feet in width.

Unconsolidated Deposits

The unconsolidated deposits in the study area consist of Wisconsinan Age glacial drift and Holocene alluvium. The glacial drift is comprised of up to 200 feet of outwash sands and gravels, and contains kames and tills. Holocene alluvium, derived from the glacial deposits, comprises the Scioto River floodplain.

The depositional history of the glacial deposits is complex. At the time of this writing, many of the interpretations of Ohio's glacial history are being reconsidered. Goldthwait (1959) believed that deposition in the area occurred in two substages, the first occurring about 50,000 years ago and the second, about 22,000 years ago.

Drilling logs (Stilson, 1976) show a layer of till or silty sand of varying thickness overlying the bedrock which is in turn overlain by heterogeneous outwash deposits. Overlying the outwash is a layer of till or alluvium. The kame complex east of the Scioto Valley which includes Spangler Hill, and the lower part of the outwash are believed to be the product of the first stage of deposition (Dreimanus and Goldthwait, 1973). The second stage of deposition is believed to have partially buried the kame complex and deposited the upper outwash and overlying till. Figure 3 shows a geologic cross section through the South Well Field.

Much of the glacial deposition was influenced by preglacial and interglacial valleys. The topography of the underlying bedrock is poorly mapped due to the lack of wells penetrating to bedrock. The bedrock surface is probably a composite surface of drainage networks developed during different glacial and interglacial stages.

The glacial deposits constitute the principal aquifer in southern Franklin County. The aquifer is unconfined with a saturated thickness of fifty-five to eighty-five feet (Norris, 1986). Eberts (1987) assigned hydraulic conductivities of 200 ft/d to 330 ft/d to the aquifer in the Scioto River Valley. Vertical hydraulic conductivities are believed to be much less. This aquifer originally discharged to the Scioto River in the study area, but now receives,in places,river water due to pumping at the South

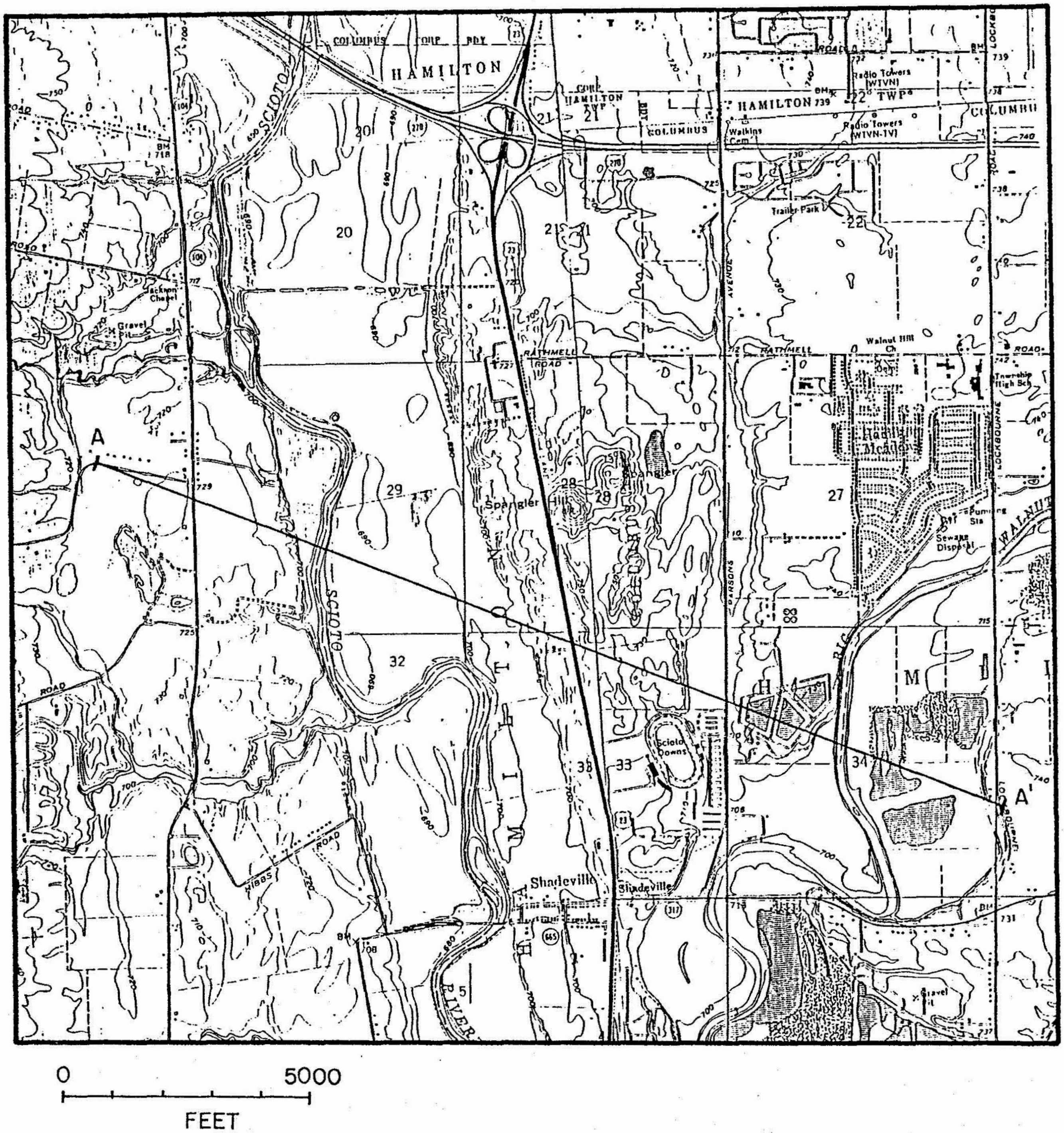


Fig. 2 Topography of Study Area and Vicinity with Location of Geologic Section A - A'

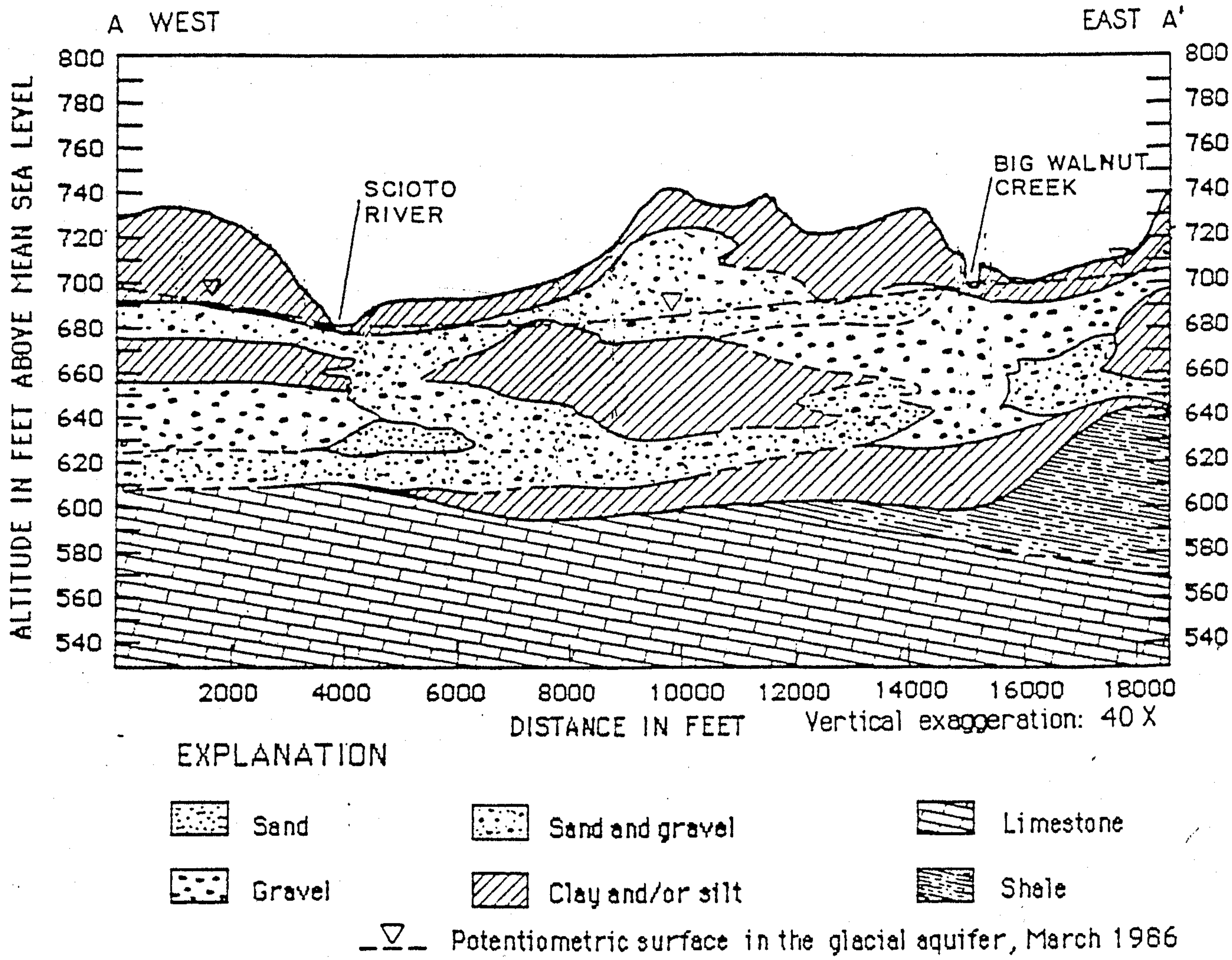


Fig. 3 Geologic Section A-A' (Modified from Eberts, 1987)

Well Field and at the American Aggregates quarry located immediately north of Interstate 270 and west of the river.

Bedrock

South-central Franklin County is underlain by the Columbus Limestone, Delaware Limestone, Olentangy Shale, and Ohio Shale. These Devonian Age formations are concealed by the Wisconsinan glacial deposits and the location of contacts can only be inferred from well logs, quarries, and regional attitude and thickness. Regional dip is approximately thirty-one feet per mile south of east (Stout, 1941). Collector Wells 101, 103, and 104, along the Scioto River, are completed just above the limestone surface, whereas Collector Well 115, along Big Walnut Creek, is completed to shale (Stilson, 1977). A generalized stratigraphy is presented in Table 2.

The Columbus and Delaware Limestones constitute a confined aquifer. Eberts (1987) assigned the bedrock aquifer a hydraulic conductivity of 15 ft/d.

Scioto River

The Scioto River is 230.8 miles long and has a drainage area of 6,517 square miles (Cross, 1967). It originates in Auglaize County near Waynesfield and empties into the Ohio River at Portsmouth. The study area is a 4.3-mile segment from Interstate 270 to State Route 665. The average

Table 2

Stratigraphy of south-central Franklin County, Ohio

(Modified from Eberts, 1987)

System	Series	Group or Formation	Maximum Thickness, ft.	Lithology
Quaternary	Holocene		20	Silt, clay and sand deposited on the flood plains of major streams, alluvium
	Pleistocene		200	Glacial outwash deposited as surficial valley trains during Wisconsinan time, sand and gravel
			250	Lenses of sand and gravel up to 50 feet thick beneath thick tills, deposited in buried valleys
			250	Till, heterogenous mixture of silt, clay and sand and gravel, scattered with lenses of sand and gravel
Devonian	Upper	Ohio	450	Blue-black to dark-brown carbonaceous shale, grading from massive to thinly bedded
		Olentangy	30	Soft, argillaceous, blue shale with some argillaceous limestone
	Middle	Delaware	35	Thinly bedded to massive blue-gray limestone and calcareous brown shales
		Columbus	105	Brown dolomitic limestone and gray calcareous crystalline limestone

gradient of the river is 2.4 feet per mile in the study area.

The U.S. Geological Survey maintains a gage (03227500) on the west bank of the Scioto River at the Jackson Pike Waste Water Treatment Plant approximately 2.75 miles north of Interstate 270. All discharge data in the study are taken from records from this gage. These data include the return flow from the Jackson Pike Plant. The average discharge of the river is 1,405 cubic feet per second (cfs). The estimated maximum discharge was 138,000 cfs during the flood of March 25, 1913. The minimum recorded discharge was 47 cfs on September 6, 1930. It is unlikely that the discharge will ever drop this low again due to the fact that approximately 130 cfs is contributed by the Jackson Pike Plant.

The Scioto River is a run dominated stream. In the study area approximately seventy percent of the river was mapped as run dominated, whereas pools and riffles accounted for eighteen and twelve percent, respectively. The streambed, in the runs and riffles sampled, is composed of poorly sorted sand, cobble, and gravel. River-bottom profiles for each site are located in Appendix A.

PREVIOUS STUDIES

Stilson (1976) determined that approximately eighty-one percent of pumpage would be derived from induced stream infiltration on the basis of a hydrologic budget which was in part based on pumping-test data. The three components of the hydrologic budget were recharge from precipitation, underflow, and average induced stream infiltration. The tributary area of the well field was considered to be at least 30,000 feet long (northward from 5,000 feet south of Shadeville) and 15,000 feet wide. This budget was also based on five radial collector wells along the Scioto River.

Two types of pumping tests were performed. A step-drawdown test was performed to determine well capacity and well efficiency. A constant-rate pumping test of three to six days duration was performed to determine transmissivity, hydraulic conductivity, storativity, and the effective distance to the line of recharge. Stilson (1976) used deep and shallow observation wells in lines perpendicular and parallel to the Scioto River. River well points were also installed four to six feet deep in the streambed.

Stilson (1976) arrived at a value for average induced infiltration in the following manner. First, the effective distance to the line of recharge was computed. This distance may be derived from the Theis nonequilibrium equation or the Thiem equation (Rorabaugh, 1956). This is based on drawdown in any two observation wells in a line. For a line of observation wells perpendicular to the river (between the pumped wells and the river), the distance is computed from the following equation.

$$\frac{s_1}{s_2} = \frac{\log\left(\frac{2a-r_1}{r_1}\right)}{\log\left(\frac{2a-r_2}{r_2}\right)}$$

For a line of observation wells parallel to the river, this distance is computed from

$$\frac{s_2}{s_1} = \frac{\log \frac{\sqrt{4a^2+r_1^2}}{r_1}}{\log \frac{\sqrt{4a^2+r_2^2}}{r_2}}$$

where

s = drawdown in observation well (ft),

r = distance from observation well to pumped well (ft), and

a = effective distance to line of recharge (ft).

Next, the percentage of water derived from the Scioto River was calculated from Theis (1941).

$$P_r = \frac{2}{\pi} \int_0^{\frac{\pi}{2}} e^{-f \sec^2 u}\, du$$

where

P_r = percentage of pumped water derived from river,

$u = \tan^{-1} (\frac{r_r}{a})$

$f = 2693\ a^2 S/Tt$

r_r = distance along line source from a perpendicular from pumping well to a (ft),

a = effective distance to line of recharge (ft),

S = storativity,

T = transmissivity (gpd/ft^2), and

t = time (days).

Stilson (1976), citing Norris (1969), stated that approximately one-half of the river water would originate from a length of river equal to twice the effective distance to the line of recharge. The infiltration rate was calculated from

$$I_r = \frac{0.5\ P_r Q}{2aw}$$

where

I_r = infiltration rate through the stream bed (gpd/ft^2),

Pr = percentage of water derived from river,

Q = pumping rate (gpd),

a = effective distance to line of recharge (ft), and

w = width of river (ft).

The infiltration rate was then standardized to a potential infiltration rate (gpd/ft^2/ft of head loss) by dividing it by the average head loss. This value was in turn multipled by a mean river depth to obtain an average infiltration rate. The yield, based on this rate, was computed to be 41.6 Mgd. Underflow and recharge from precipitation account for an additional 9.8 Mgd, for a total hydrologic budget of 51.4 Mgd.

Norris (1986) evaluated the Stilson (1976) pumping-test data and determined that Stilson had made some interpretative errors. Furthermore, Norris (1986) suggested that the hydraulic connection between the river and the aquifer is poor.

Norris (1986) noted that pre-pumping ground-water levels were up to four feet higher than river level. This head difference was evident in both deep and shallow wells leading Norris to believe that low permeability materials were found throughout the aquifer.

Norris (1986) felt that drawdowns in river well points were actually the result of radial flow to the wells and not vertical flow through the streambed. Well points driven four to six feet into the streambed would, in places, penetrate the top of the aquifer. An error in interpreting drawdown would affect the calculation to determine the distance to the effective line of recharge. Additionally, Norris (1986) found the distances to the effective lines of recharge to be excessive. These distances should approximately coincide with the distance to the river, but were instead up to 4,000 feet beyond it.

Transmissivity values computed by Norris (1986) were lower than those computed by Stilson (1976). Transmissivity, along with distance to the effective line of recharge, are variables used to determine the percentage of pumpage derived from the river.

Finally, volume-of-cone analyses performed by Norris (1986) indicated a poor hydraulic connection. In this method, the volume of the cone of depression is related to the amount of water pumped to determine a storativity value. Storativity values for all sites except 106 (south of Shadeville) were in the confined range. These values were compared to values obtained for wells along the Scioto River at Piketon (Norris, 1969). The Piketon data indicated storativity values greater than unity, suggesting that the river was recharging the aquifer.

Other studies performed at the South Well Field have included a hydrologic budget based, in part, on the Stilson (1976) data (Stowe, 1979), a stream-aquifer water-quality study (de Roche and Razem, 1984), and computer modeling studies (Weiss and Razem, 1980; Razem, 1983; de Roche, 1985; Eberts, 1987).

METHODS OF STUDY

Mapping

The first step in this field investigation was to map the Scioto River and to subdivide it into three river environments: pools, runs, and riffles. Mapping was accomplished by canoeing the study area and color coding the appropriate U.S.Geological Survey 7 1/2 minute quadrangles. The data were then plotted on a map with an exaggerated river width (Figure 4).

Definitions of river environments tend to be more qualitative than quantitative, and vary among workers. Geomorphologists and sedimentologists use a classification based solely on pools and riffles, and do not recognize runs. For the purposes of this study, the terms pool, riffle, and run are defined as follows. A riffle is a shallow portion of a stream, generally one foot or less deep, where the stream surface is broken by turbulence due to streambed composition. A pool is defined as that part of a stream with little discernable current or a topographically depressed bottom. Major pool sections were generally more than four feet deep. Other portions that qualify as pools were backwater areas behind bars or islands. These areas, however, were too small to map. A run is defined as an area of fairly uniform cross section with a discernable current. Commonly, there was no clear-cut

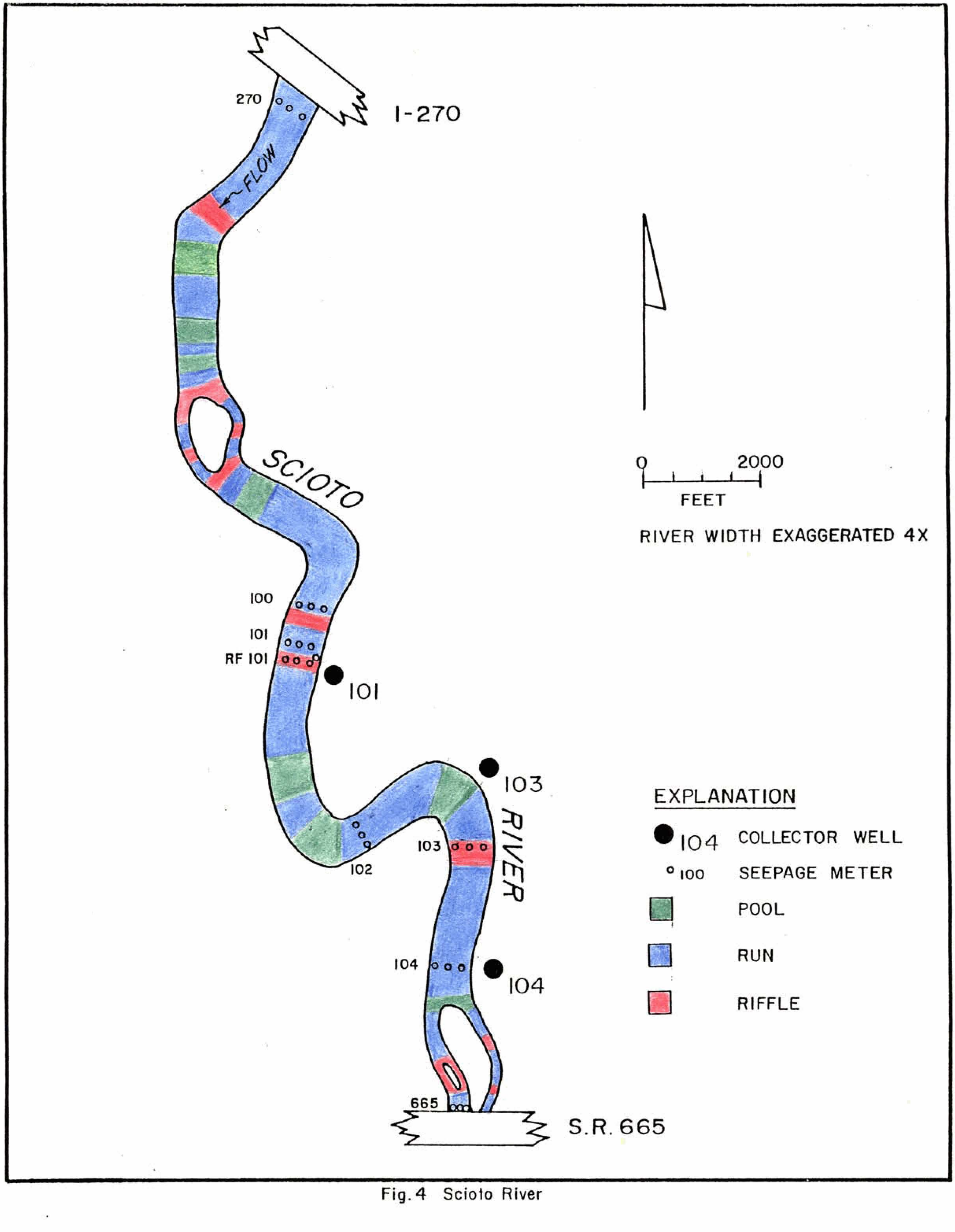


Fig. 4 Scioto River

distinction as to where a run ended and a pool began. Additionally, these classifications will vary with river stage and with time. The river, in this study, was mapped during low-flow conditions.

Streambed Vertical Hydraulic Conductivity Measurements

Vertical hydraulic conductivity measurements were determined by using a seepage meter and a piezometer. The seepage meter was used to measure Darcy velocity and the piezometer was used to measure vertical hydraulic gradient. Vertical hydraulic conductivity then was determined from Darcy's law, which is rewritten in terms of hydraulic conductivity.

$$K = \frac{Q}{A(dh/dl)}$$

where

K = vertical hydraulic conductivity,

Q = flow across the portion of streambed covered by the seepage meter,

A = cross-sectional area of the seepage meter, and

dh/dl = vertical hydraulic gradient.

The seepage meter and piezometer were placed one to two feet apart. At seven sites along the Scioto River, three sets of seepage meters and piezometers were placed in a line across the river (perpendicular to the banks). At an eighth site, four stations were used. River-bottom profiles were measured at each site with a tape and level rod.

The profiles and river stages for each site are included in Appendix A.

Seepage meters were constructed from both ends of a steel 55-gallon drum. Handles were welded on top of the instruments. A two-inch diameter rubber stopper is inserted into the top or side of the drum and a tube passing through the stopper on one end is attached to a plastic bag on the other end (see Figures 5 and 6). Cherry and Lee (1978) provide a detailed description of seepage meter construction. Modifications to the manner described by Cherry and Lee (1978) included replacing the flexible plastic tubing with a length of larger half inch diameter surgical tubing and using a shorter six inch high barrel for shallow riffles.

Piezometers were constructed from five to six foot lengths of 1.5-inch diameter steel pipe. Both ends of the pipe were threaded. A sand point with a six-inch screen was attached to one end and a removable cap was attached to the other end.

Choosing a suitable location to place the instruments is imperative. Water depth had to be greater than approximately six inches so that the seepage meters could be submerged entirely. Water deeper than approximately two and a half feet was too difficult to work in. On the Scioto River, these methods are suitable only under low-flow conditions of late summer and early fall.

Fig. 5 Seepage Meter in Streambed

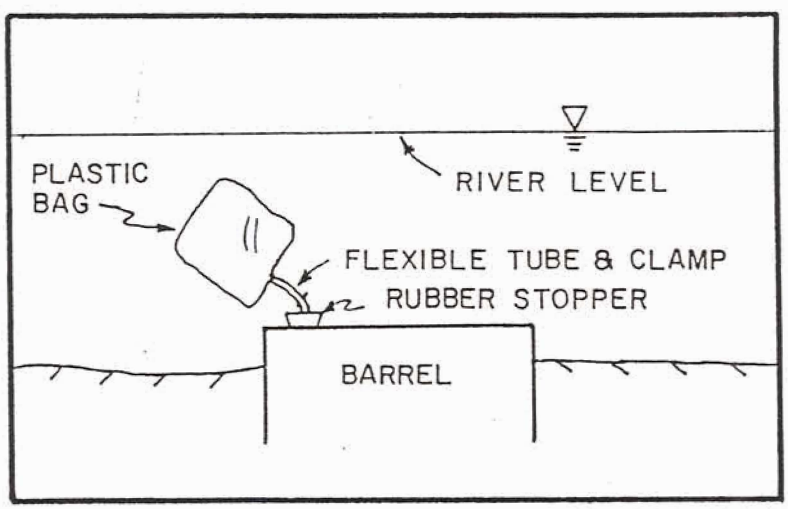


Fig. 6 Side View of Seepage Meter in Streambed

Measurements were not attempted if the discharge at the U.S. Geological Survey gage (03227500) was greater than 300 cfs.

Because of the gravel and cobble character of the Scioto River streambed, placement of the seeepage meter commonly was laborious. The instruments were rotated into the streambed as far as possible, then pounded in with a steel post driver. All but three or four inches of the barrel were driven into the bottom to create an effective seal. With the shorter riffle version, all but approximately one or two inches of the barrel were submerged below the streambed. The seepage meters generally were allowed to remain undisturbed in the streambed for 24 hours after emplacement and before measurements were made. This allowed shifting silt and sand to settle around the bottom of the drum creating a tighter seal.

Seepage-meter measurements were taken in the following manner. The plastic bag was filled with 500 ml of water. After squeezing air from the bag, the flexible rubber tube was clamped and the stopper was inserted into the opening on the barrel. Next, the clamp was removed. After waiting ten to twenty minutes (timed with a stop watch), the tube was clamped and the bag was removed. The volume of water was then measured with a graduated cylinder and the change in volume was recorded. The change in volume divided by the time yields the flow rate Q. The flow rate divided by the cross sectional area of the seepage meter yields a Darcy velocity. A minimum of three measurements were taken at each station and a mean Darcy velocity was calculated at each station.

Piezometers were driven eighteen inches into the stream bed (as measured from from the center of the screen). Water-level measurements were taken with a steel tape in the following manner. First, the distance from the top of the piezometer to the water level in the river was measured. Second, the distance from the top of the piezometer to the water level inside the pipe was measured. This was accomplished by rubbing chalk on the first foot of the tape and subtracting the water line reading from the reading at the top of the piezometer. The head difference is the difference between the river level and the potentiometric surface inside the pipe (Figure 7). The head difference divided by the distance the piezometer is driven into the streambed, yields the hydraulic gradient.

If the head difference in a piezometer was less than 0.2 feet, it was checked against a measurement made with a manometer. In a few cases, the potentiometric surface could not be reached at eighteen inches and the piezometer was driven into the streambed an additional amount, usually six to twelve inches. All piezometers were developed by pouring water down the pipe and agitating with an aluminum rod.

Sediment Samples

Sediment samples were taken at each station by inverting a plastic cup and twisting it into the streambed (Figure 8). A hole was scooped out by hand next to the cup and a hand was placed over the cup as it was brought to the surface. The sample was then emptied into a plastic bag. All samples were oven-dryed and sieved using standard U.S. Geological Survey sieving procedures. Cumulative curves plotted for grain-size distribution analysis are located in Appendix B.

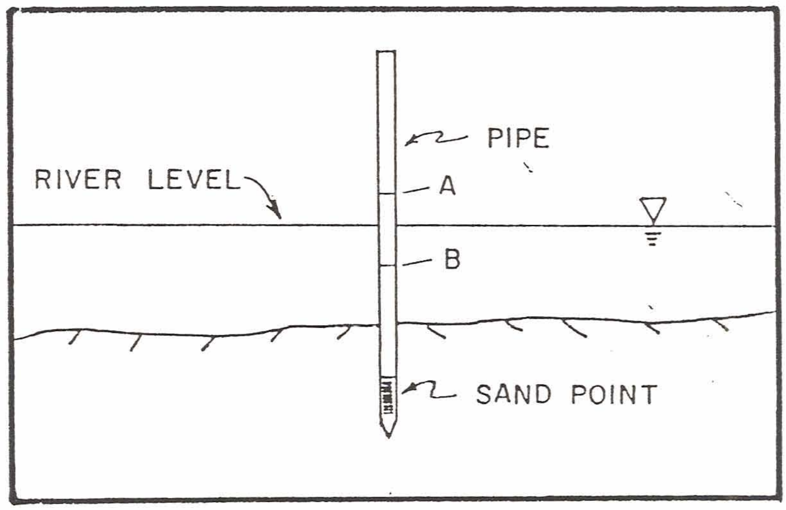


Fig. 7 Piezometer in Streambed
A - Water Level in Gaining Stream
B - Water Level in Losing Stream

Fig. 8 Sediment Sample Collection

DATA ANALYSIS AND INTERPRETATION

Hydraulic conductivity is a measure of the ability of a geologic material to transmit water and may vary spatially or with direction. Vertical hydraulic conductivity is commonly lower than horizontal hydraulic conductivity. In a vertical flow path, several layers with widely varying hydraulic conductivities may be encountered. Freeze and Cherry (1979) state that a formation with layered heterogeneity can be replaced by a single homogenous anisotropic layer. This is described by the following equation.

$$K_v = \frac{d}{\sum_{i=1}^{n} d_i/K_i}$$

where

K_v = equivalent vertical hydraulic conductivity,

K_i = vertical hydraulic conductivity of an individual layer,

d_i = thickness of an individual layer, and

d = total thickness.

Hydraulic conductivity is not generally considered a temporally variable property, however, in the case of induced stream infiltration temporal variations may be an important consideration. Hydraulic conductivity is a function of the properties of the fluid as well as the properties of the media. One of the important fluid properties is viscosity, which is a function of temperature. River water temperatures may vary approximately 30°C seasonally, whereas seasonal variations in ground-water temperature are negligible. Hydraulic conductivity values increase approximately 2.7 percent for every 1°C increase in temperature. Another source of temporal variation is the nature of the river itself. Rivers are dynamic systems in which sediment is continually being transported and deposited.

Seepage-Meter Data

Vertical hydraulic conductivity measurements were taken at twenty-five stations at eight sites. Six sites were located in run segments and two sites in riffle segments. Additionally, one measurement was taken in a shallow backwater pool. Difficulties were encountered at four of the twenty-five stations, reducing the number of data points to twenty-one. These difficulties included inability to seal the seepage meter from the river (twice) and inability to measure the head difference in the piezometer (twice).

Vertical hydraulic conductivity data were adjusted for temperature using the following equation modified from Walton (1970).

$$K_v(ADJ) = K_v \frac{\mu}{\mu_T}$$

where

$K_V(ADJ)$ = vertical hydraulic conductivity of the stream bed standardized to 12°C (ft/d),

K_V = vertical hydraulic conductivity of the stream bed at test temperature (ft/d),

μ = Coefficient of dynamic viscosity at river water temperature during test (cgs units), and

μ_T = Coefficient of dynamic viscosity at 12°C (cgs units).

Coefficients of dynamic viscosity were taken from a graph in Walton (1970) (Figure 9).

Adjusted vertical hydraulic conductivity data are listed in Table 3. Values ranged from a low of 0.04 ft/d to a high of 3.87 ft/d with a mean of 0.90 ft/d and a median of 0.31 ft/d. These values correspond to the silt or silty sand categories listed in Freeze and Cherry (1979) (Table 4).

Norris (1983) determined vertical hydraulic conductivities data from the Scioto River stream-bed at Piketon, Ohio, using pumping-test data. The mean value at Piketon was 7.5 ft/d, or about an order of magnitude higher than the South Well Field values.

Vertical hydraulic conductivity values at the South Well Field, though apparently low, still range over two orders of magnitude. The areal distribution of the data are plotted in Figure 10. This variability demonstrates the heterogeneity of the streambed. The lowest values encountered were at the collector well sites. The highest site mean value was at State Route 665.

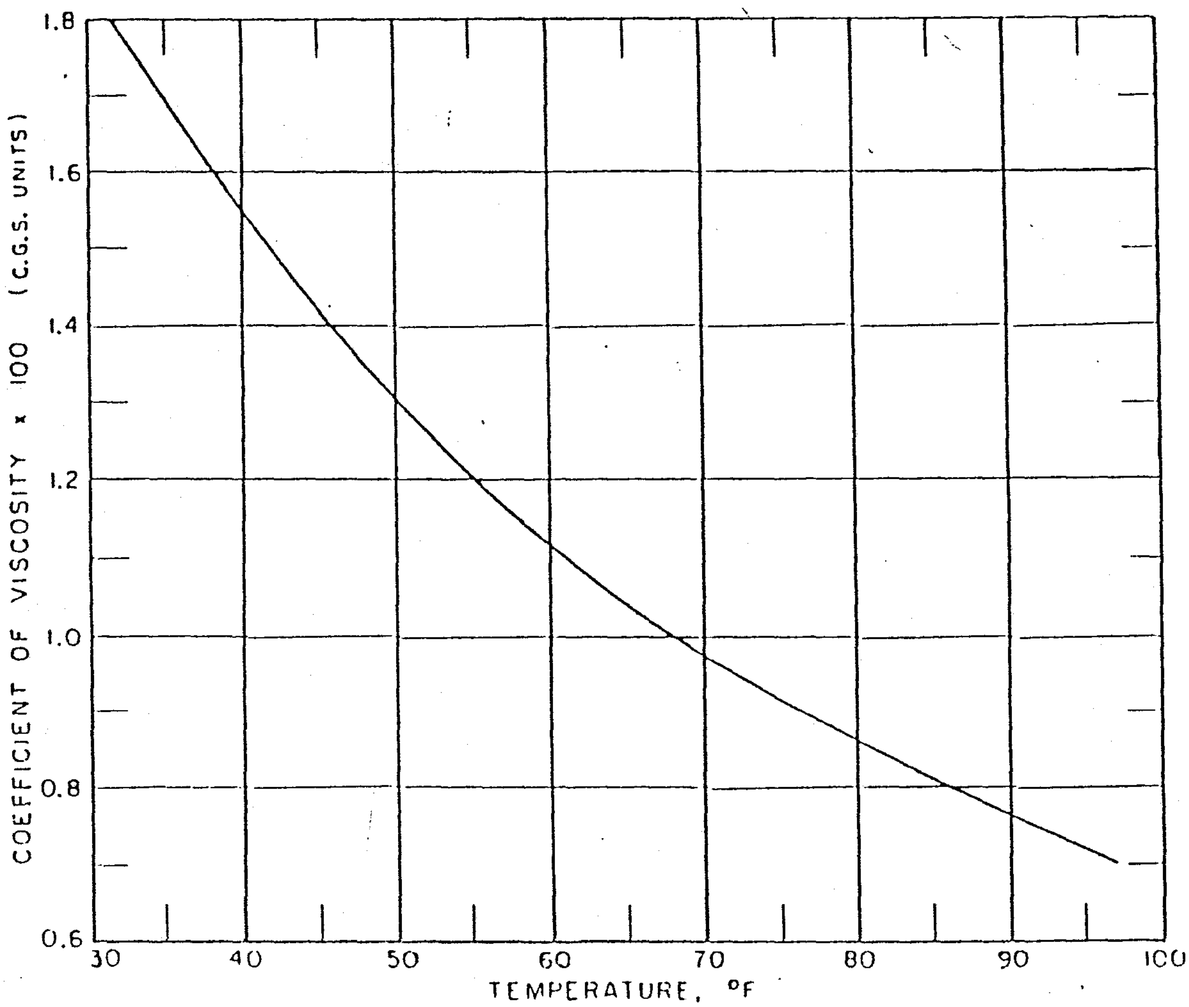


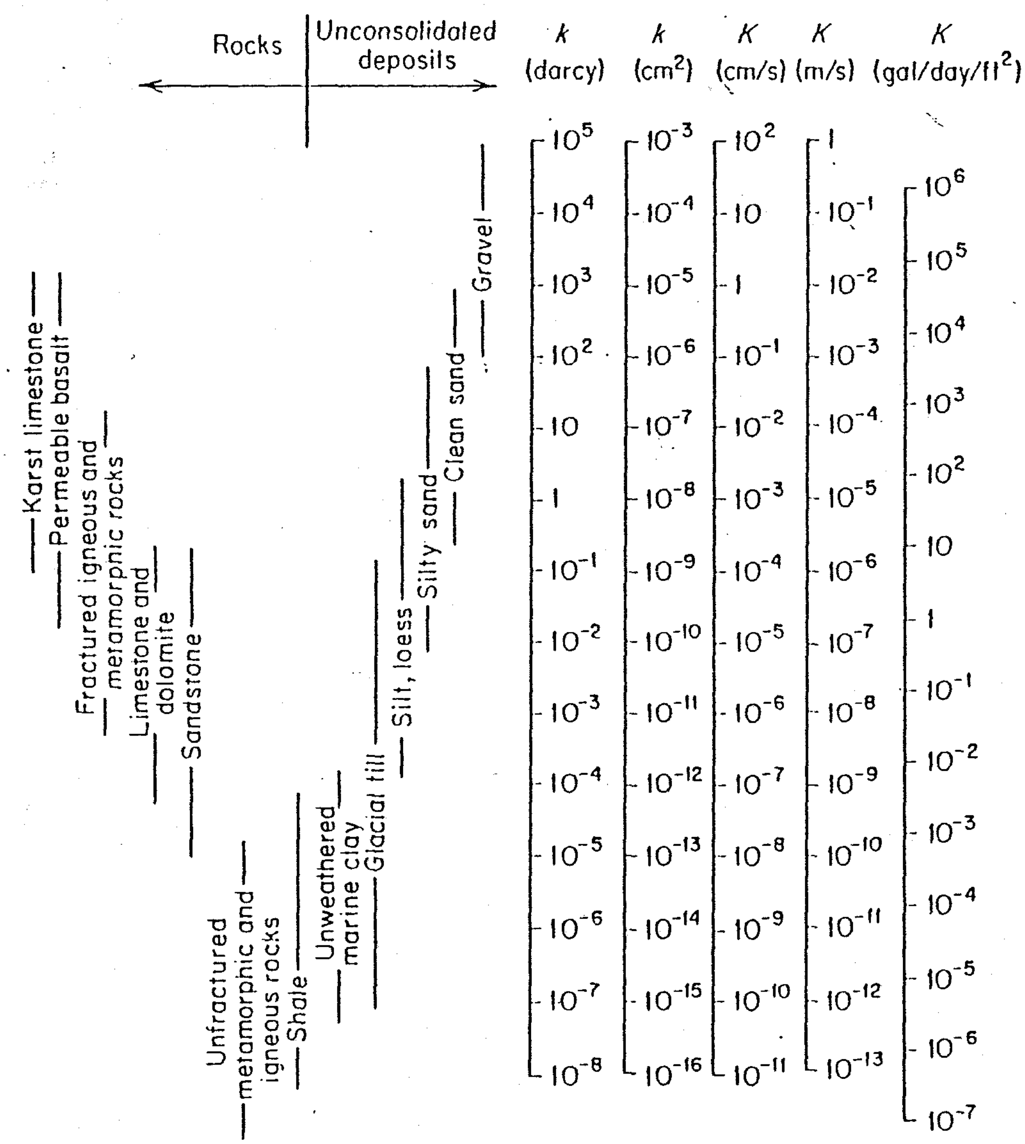
Fig. 9 Relationship between coefficient of viscosity and temperature. (*Drawn by W. C. Walton.*)
(From Walton, 1970)

Table 3
Vertical Hydraulic Conductivity Values

Station	Riverine Setting	Unadjusted K_V (ft/d)	Temp °C	K_V adjusted to 12.0° (ft/day)
104 A	Run	0.08	22.5	0.06
104 B	Run	0.08	22.5	0.06
104 C	Run	--	22.5	--
270 A	Run	0.71	23.0	0.54
270 B	Run	0.15	23.0	0.11
270 C	Run	0.11	23.0	0.08
101 A	Run	0.64	22.0	0.50
101 B	Run	0.09	22.0	0.07
101 C	Run	0.78	22.0	0.61
RF101 A	Riffle	3.02	22.0	2.36
RF101 B	Riffle	0.26	22.0	0.20
RF101 C	Riffle	0.05	22.0	0.04
RF101 D	Pool	0.44	22.0	0.34
102 A	Run	4.35	17.0	3.82
102 B	Run	--	17.0	--
102 C	Run	1.08	17.0	0.95
103 A	Riffle	0.04	12.5	0.04
103 B	Riffle	--	12.5	--
103 C	Riffle	0.17	12.5	0.17
100 A	Run	--	14.5	--
100 B	Run	0.33	14.5	0.31
100 C	Run	0.19	14.5	0.18
665 A	Run	1.56	11.5	1.59
665 B	Run	3.81	11.5	3.87
665 C	Run	2.99	11.5	3.04
Mean		1.00	18.3	0.90

-- unable to measure K_V because of inability to seal seepage meter or locate potentiometric surface with piezometer

Table 4 Range of Values of Hydraulic Conductivity and Permeability (From Freeze & Cherry, 1979)



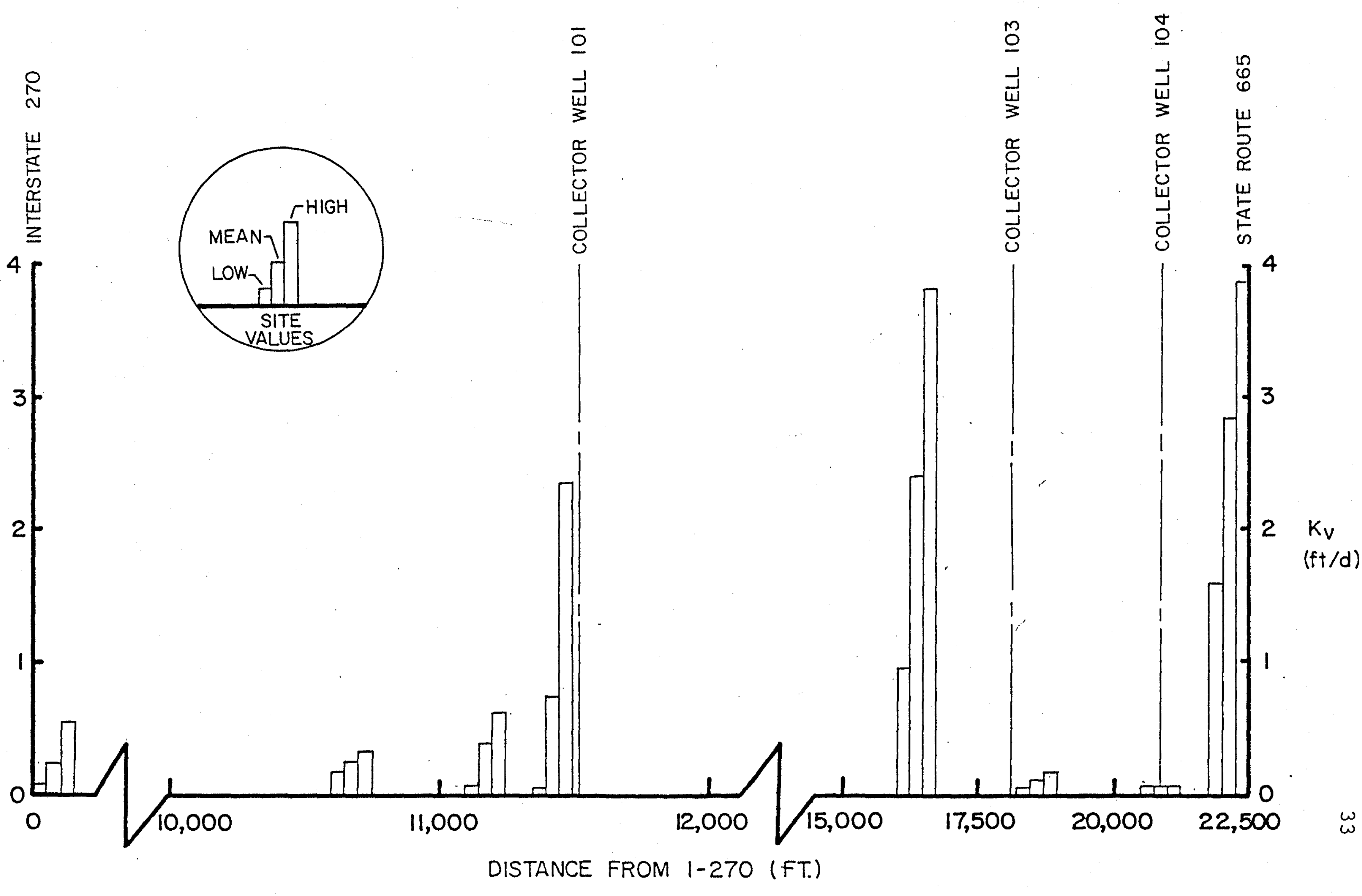


Fig. 10 Areal Distribution of Streambed Vertical Hydraulic Conductivity

Stilson (1976) reported streambed permeability in terms of infiltration rate per foot of head difference. This is equivalent to the vertical hydraulic conductivity divided by the distance from the top of the streambed to the center of the piezometer sand point. Hydraulic conductivity values from this study were converted to potential infiltration rates for comparison with data presented by Stilson (1976) (Tables 5 and 6). Fifteen out of twenty-one values were lower than values listed in Stilson (1976). Six values, however, were substantially higher. It should be noted that the Stilson (1976) values were based on the Rorabaugh (1956) method from head losses in river well points driven four to six feet into the streambed and possibly into the aquifer. Norris (1986) believed that these head losses were actually due to radial flow to the well and not vertical flow through the streambed. The results of this comparison are inconclusive.

Low streambed permeabilities may be due, in part, to the streambed being clogged by particulate from waste water plant effluent. Under low-flow consitions, over ninety percent of the water in the Scioto River can be effluent from the Jackson Pike Waste Water Treatment Plant. Eberts (1987) used a streambed vertical hydraulic conductivity of 0.22 ft/d for the Scioto River downstream from the plant and a higher value of 0.67 ft/d upstream from the plant. River discharge, however, is sufficiently high, even under

Table 5

Vertical Hydraulic Conductivity
Converted to Infiltration Rate

Station	dh/dl	Adj. 12°C K_v (ft/day)	K_v (gpd/ft^2)	dL	I_r = K dh/dl I_r (measured) (gpd/ft^2)	I_r = k/dl I_r (potential) (gpd/ft^2/ft head)
104 A	0.89	0.06	0.45	1.5	0.40	0.30
104 B	1.77	0.06	0.45	1.5	0.80	0.30
104 C	--	--	--	1.82	--	--
270 A	0.12	0.54	3.41	1.5	0.41	2.27
270 B	0.65	0.11	0.82	1.5	0.53	0.55
270 C	1.29	0.08	0.60	1.5	0.77	0.40
101 A	0.51	0.50	3.74	1.7	1.91	2.20
101 B	1.97	0.07	0.52	2.2	1.02	0.24
101 C	0.85	0.61	4.56	1.7	3.88	2.68
RF101A	0.09	2.36	17.65	1.5	1.59	11.77
RF101B	0.53	0.20	1.50	1.5	0.80	1.00
RF101C	1.14	0.04	0.30	2.7	0.34	0.88
RF101D	0.06	0.34	2.54	1.5	0.15	1.69
102 A	0.04	3.82	28.58	1.5	1.14	19.05
102 B	0.05	--	--	1.5	--	--
102 C	0.67	0.95	7.11	1.5	4.76	4.74
103 A	0.76	0.04	0.30	1.5	0.23	0.20
103 B	0.08	--	--	1.5	--	--
103 C	1.23	0.17	1.27	1.5	1.56	0.85
100 A	--	--	--	2.3	--	--
100 B	0.98	0.31	2.32	1.5	2.27	1.55
100 C	0.96	0.18	1.35	1.5	1.30	0.90
665 A	0.21	1.59	11.89	1.5	2.50	7.93
665 B	0.14	3.87	28.95	1.5	4.05	19.30
665 C	0.17	3.04	22.74	1.5	3.87	15.16
Mean		0.90	6.71		1.63	4.47

Table 6

Infiltration Rates from Stilson (1976)

Site	Pr	dh	I_r (gpd/ft^2)	Potential I_r (gpd/ft^2 head)
100	64%	0.64	1.61	2.52
101	25%	0.10	0.35	3.50
103	63%	--	0.67	--
104	--	--	--	--
104 R	59%	0.44	1.06	2.41
106	63%	0.48	1.37	2.85
Mean	55%	0.42	1.01	2.82

low-flow conditions, to provide a velocity high enough to maintain a sand, gravel, and cobble bed in most of the river.

Another factor that may influence streambed permeability is the presence of a layer of alluvium underneath the streambed. Drilling logs (Stilson, 1976) show the presence of a clay or silty clay layer at all sites except at Collector Well 103. This layer is varying in thickness. The well logs and hydraulic conductivity values determined by Norris (1986) indicate that the aquifer is very heterogenous and stratified. Assigning a single, equivalent homogeneous anisotropic streambed permeability value to the heterogeneous stream-aquifer system could conceivably lead to a vertical hydraulic conductivity even lower than that of the streambed. This could occur if a layer of very low hydraulic conductivity were to occur beneath the streambed and were incorporated into the calculation for equivalent vertical hydraulic conductivity described by Freeze and Cherry (1979).

Grain-Size Distribution Data

Sieve analyses were performed for all sediment samples and semi-logarithmic plots of weight percent finer versus grain size were made. Table 7 lists the uniformity coefficient (d_{60}/d_{10}) of each plot. There appears to be no correlation between vertical hydraulic conductivity and

Table 7

Grain Size Distribution Data

Station	Seepage Meter K_v (ft/d)	Cu (d_{60}/d_{10})	d_{10} (mm)	d_{10}^2	Hazen Approximation K_v (ft/d)
104A	0.06	24.1	0.79	0.62	176
104B	0.06	12.9	1.75	3.06	867
104C	–	–	–	–	–
270A	0.54	19.4	0.62	0.38	108
270B	0.11	26.5	0.49	0.24	68
270C	0.08	13.0	0.74	0.55	156
101A	0.50	52.1	0.48	0.23	65
101B	0.07	22.1	0.95	0.90	255
101C	0.61	3.25	8.00	64.0	18,142
RF101A	2.36	21.3	0.46	0.21	60
RF101B	0.20	26.8	0.71	0.50	142
RF101C	0.04	27.9	1.40	1.96	556
RF101D	0.34	–	–	–	–
102A	3.82	2.67	6.00	36.0	10,205
102B	–	–	–	–	–
102C	0.95	6.80	5.00	25.0	7,087
103A	0.04	14.6	2.40	5.76	1,633
103B	–	–	–	–	–
103C	0.17	20.5	1.10	1.21	343
100A	–	–	–	–	–
100B	0.31	26.2	0.42	0.18	51
100C	0.18	9.50	4.00	16.00	4,535
665A	1.59	15.6	0.39	0.15	43
665B	3.87	8.33	0.30	0.09	26
665C	3.04	37.5	0.24	0.06	17

uniformity coefficients. The Hazen equation was used to estimate hydraulic conductivity values.

$$K = d_{10}^{2}$$

where

K = hydraulic conductivity (cm/s),

d_{10} = grain size diameter where 10 percent of grains are finer (mm).

This value was then lowered by one order of magnitude to approximate vertical hydraulic conductivity. Again there was no correlation with seepage-meter data (Table 7). It appears that the Hazen equation is not valid for poorly sorted sediments. Norris (1988) stated that grain size distribution data from Piketon were inconclusive also. Cumulative-grain size curves are included in Appendix B.

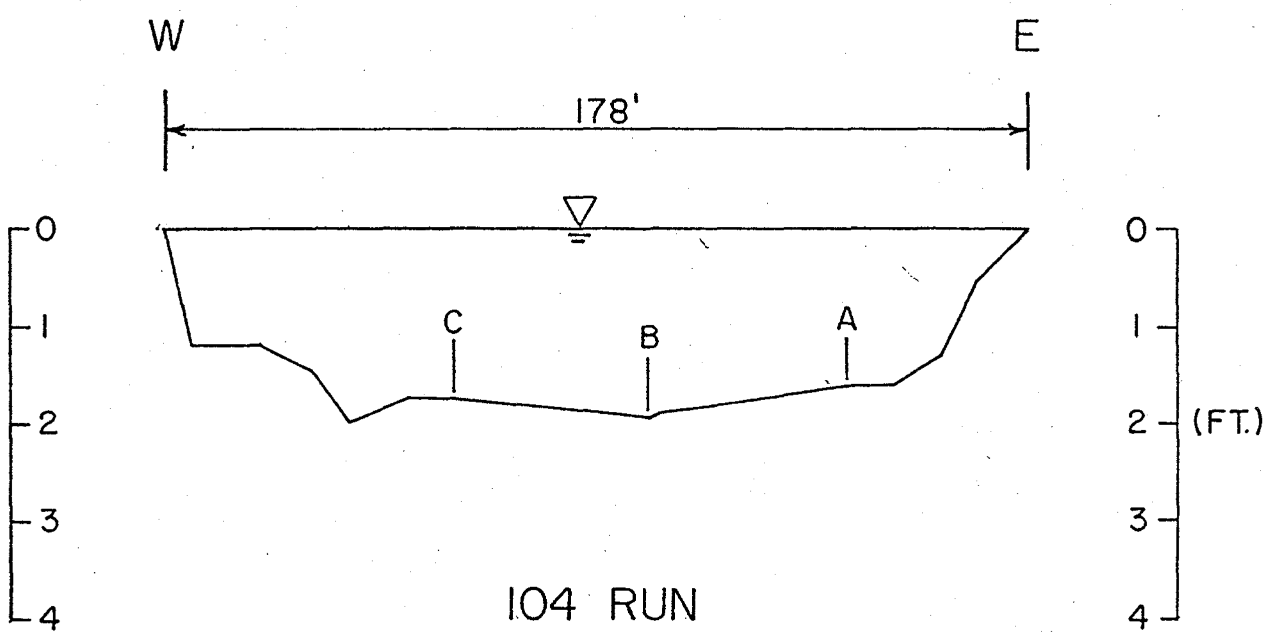
CONCLUSIONS

The following conclusions were reached for this study:

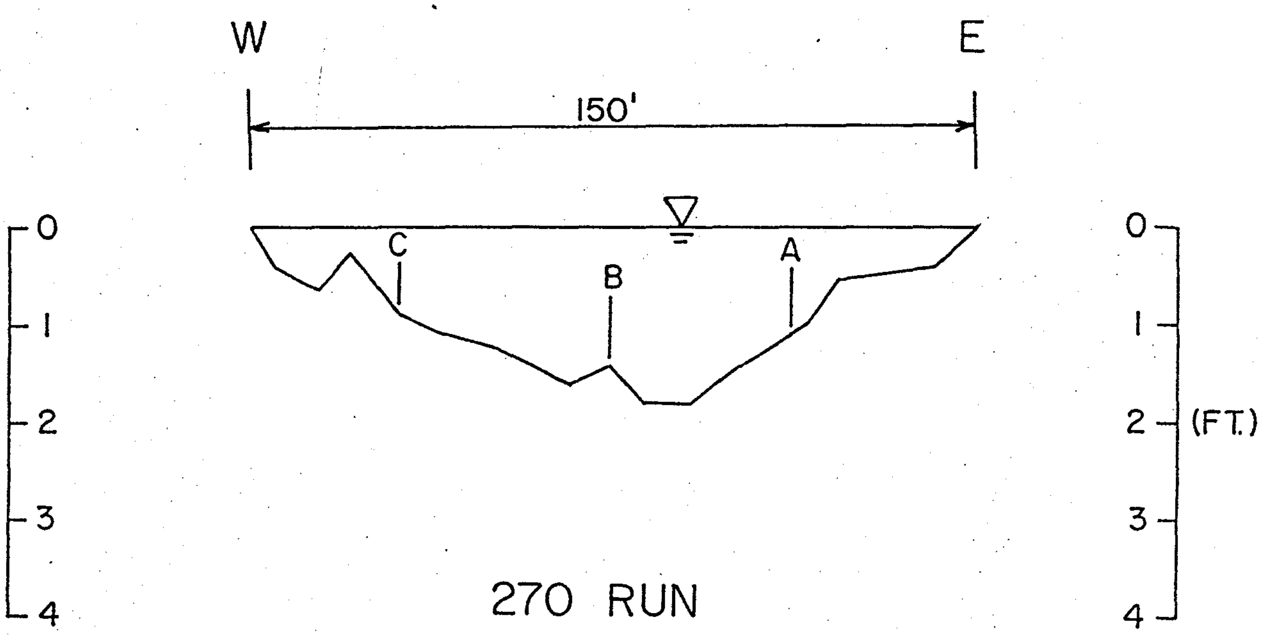
- Streambed vertical hydraulic conductivities are low. Values obtained from direct measurement of the streambed with a seepage meter and piezometer are two to four orders of magnitude lower than hydraulic conductivities of the aquifer.
- The streambed is heterogeneous with regard to vertical hydraulic conductivity distribution. It would be difficult to assign a single value to the entire streambed. The lowest values, however, were measured along stream segments adjacent to the collector wells.
- Alluvium underlying the streambed and aquifer stratification may contribute to an overall low vertical hydraulic conductivity of the stream-aquifer system. This value may be lower than the streambed itself.
- Comparison with streambed infiltration rates by Stilson (1976) are inconclusive, however Norris (1986), in a reevaluation of the Stilson (1976) data, felt that the Stilson values for streambed permeability and the percentage of water derived from the Scioto River were excessive.
- Analysis of grain-size distribution was inconclusive.

APPENDIX A

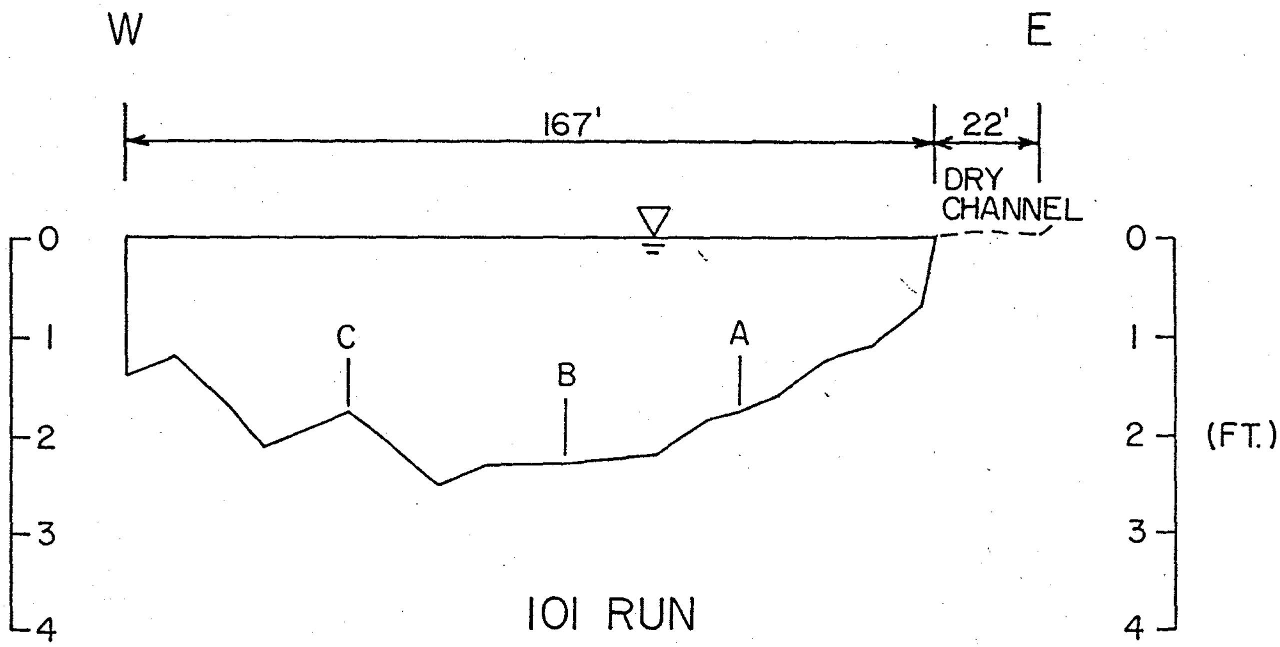
RIVER PROFILES



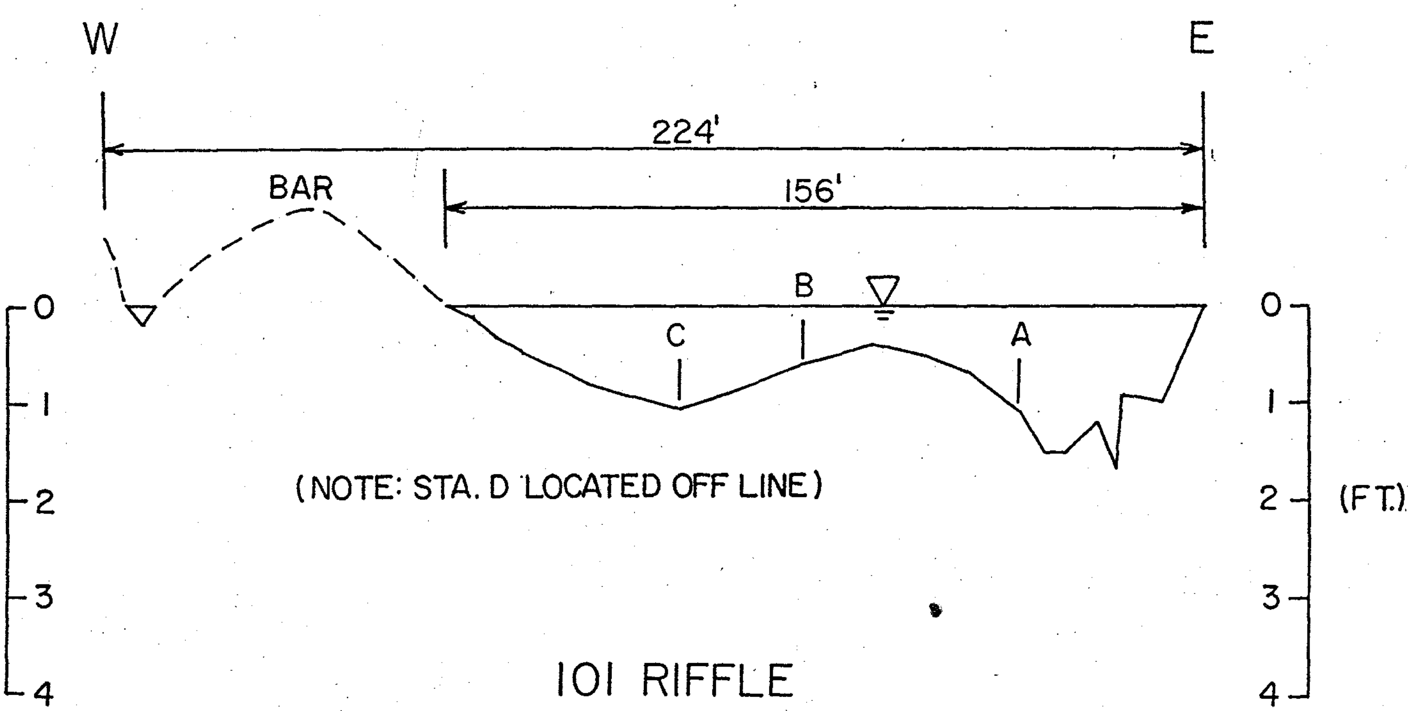
104 RUN
9-7-88
237 c.f.s.



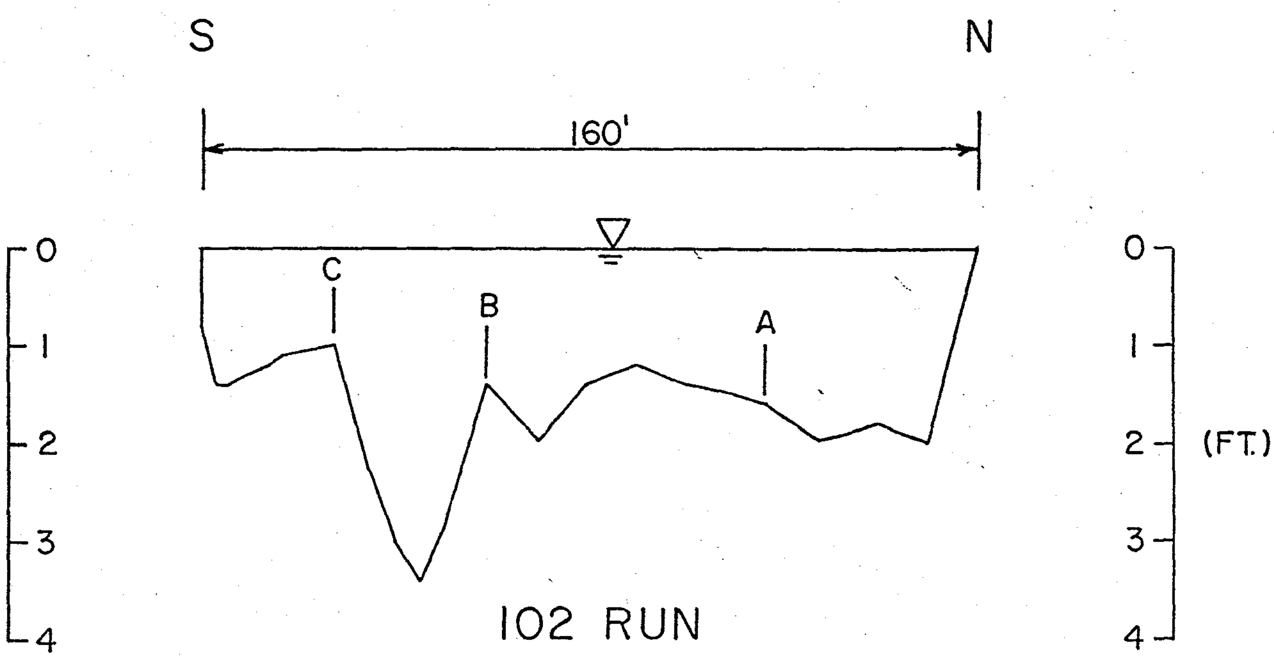
270 RUN
9-15-88
193 c.f.s.



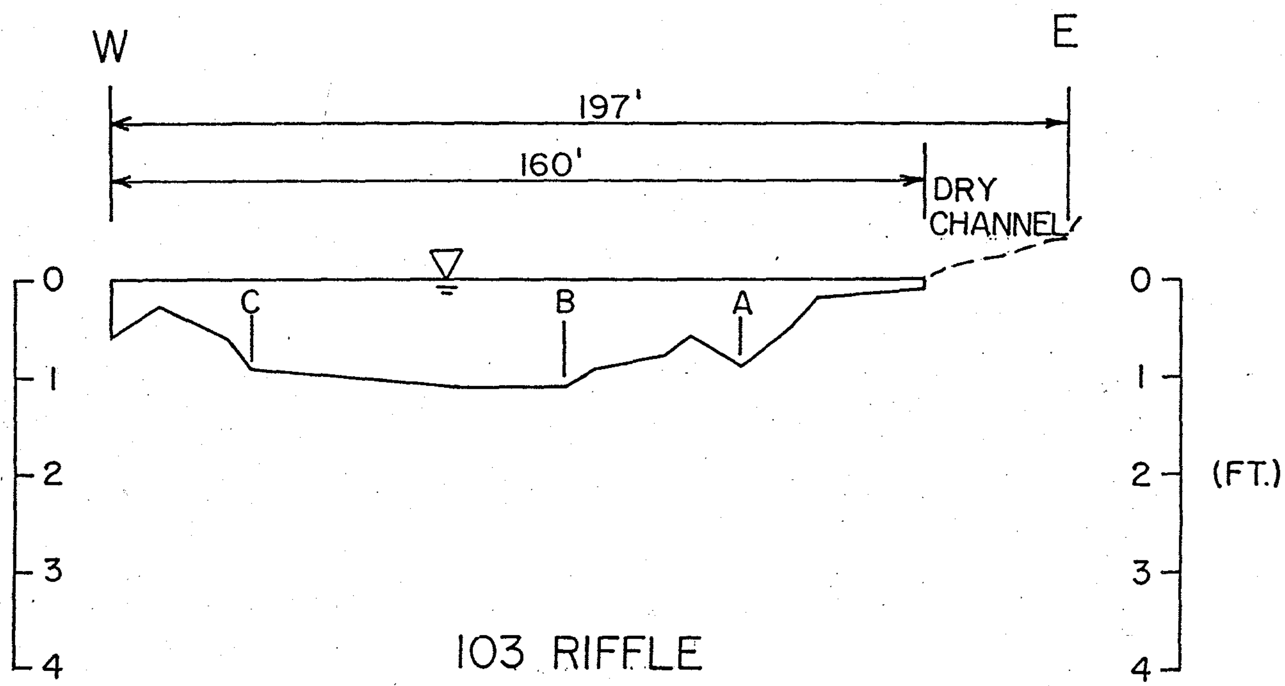
101 RUN
9-22-88
267 c.f.s.



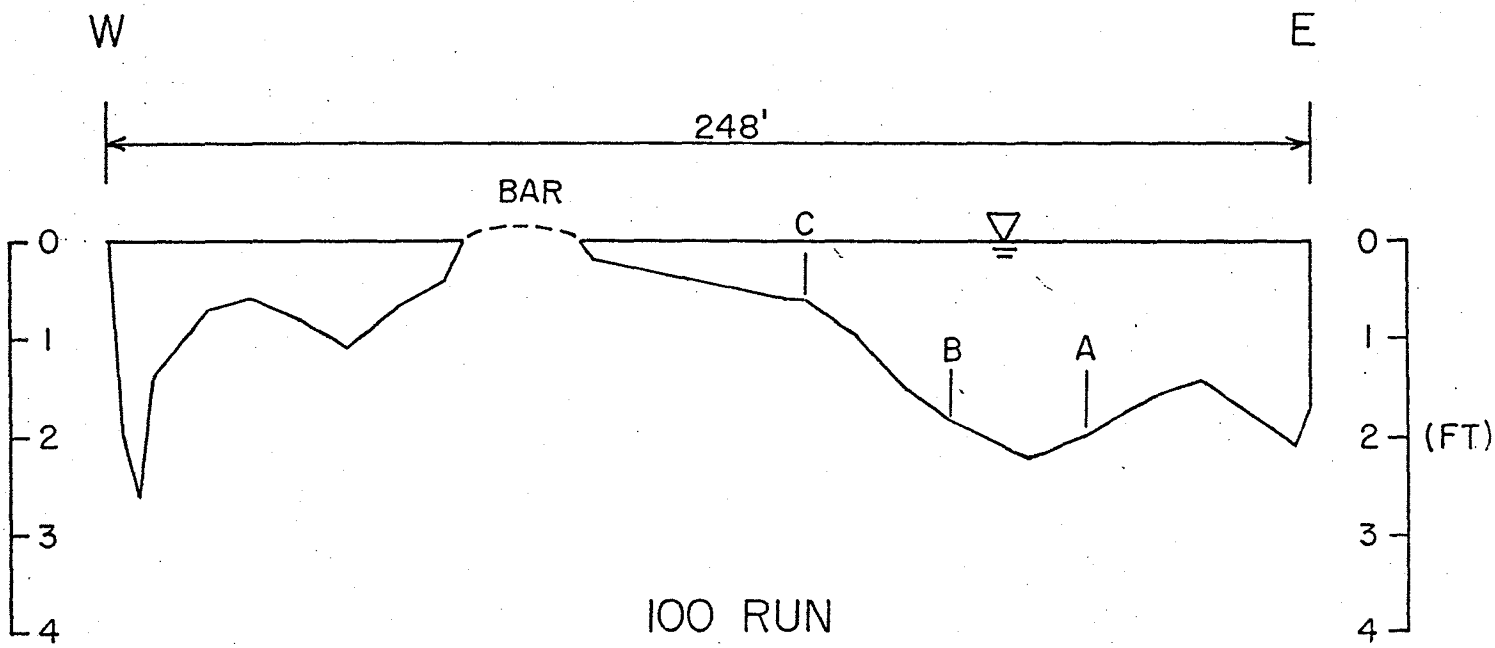
101 RIFFLE
9-28-88
146 c.f.s.



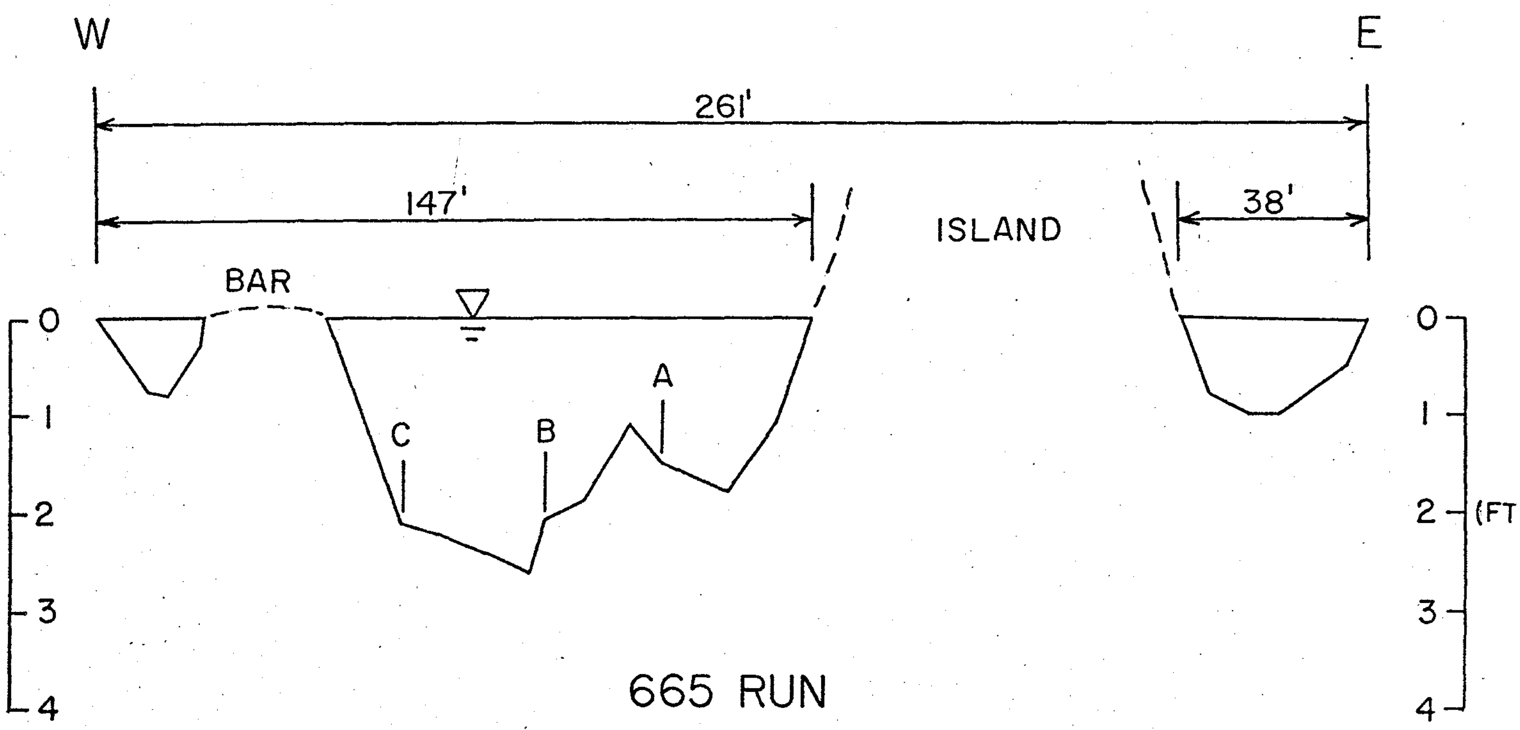
102 RUN
10-5-88
144 c.f.s.



103 RIFFLE
10-13-88
141 c.f.s.



100 RUN
10-19-88
177 c.f.s.

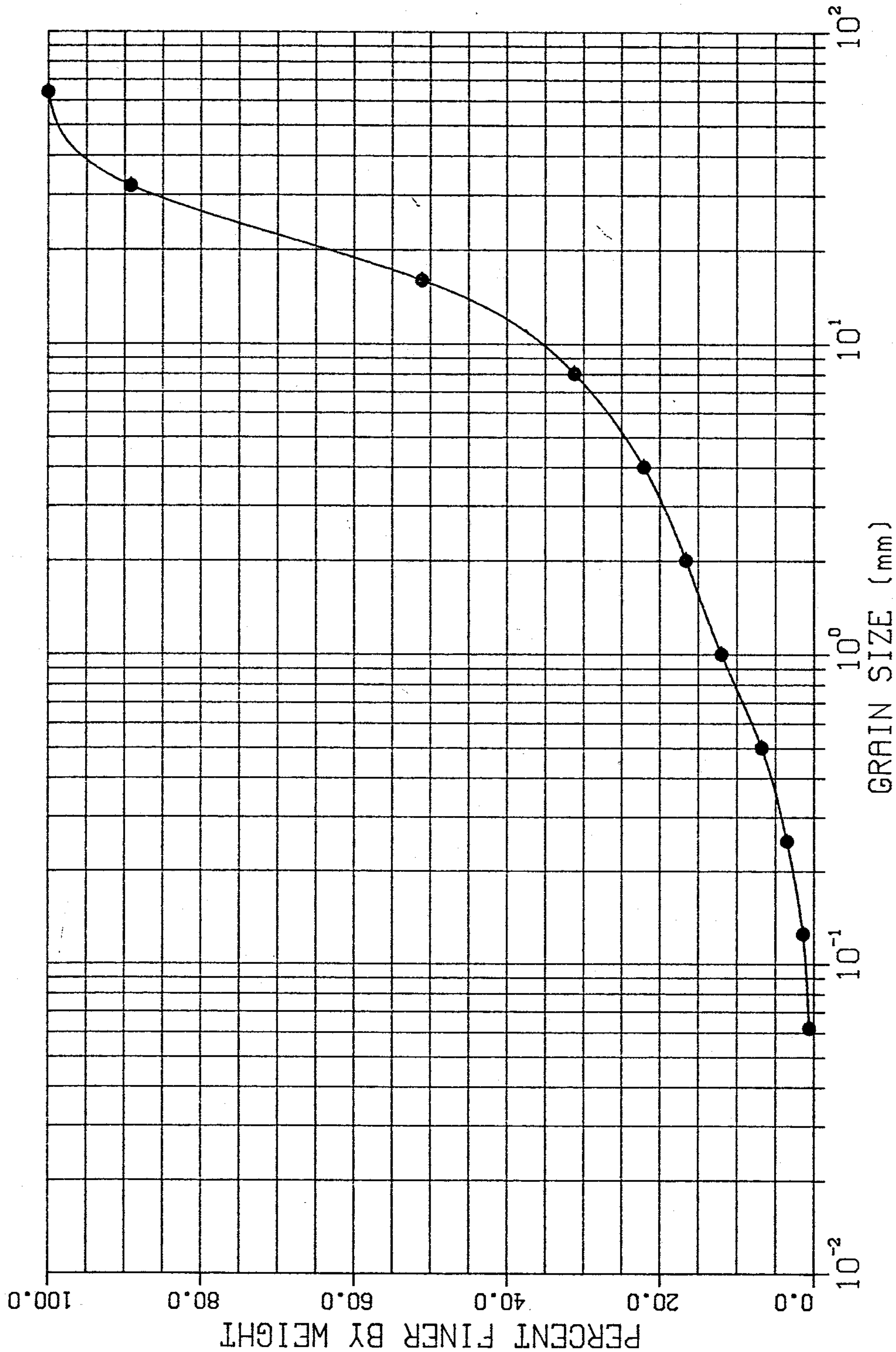


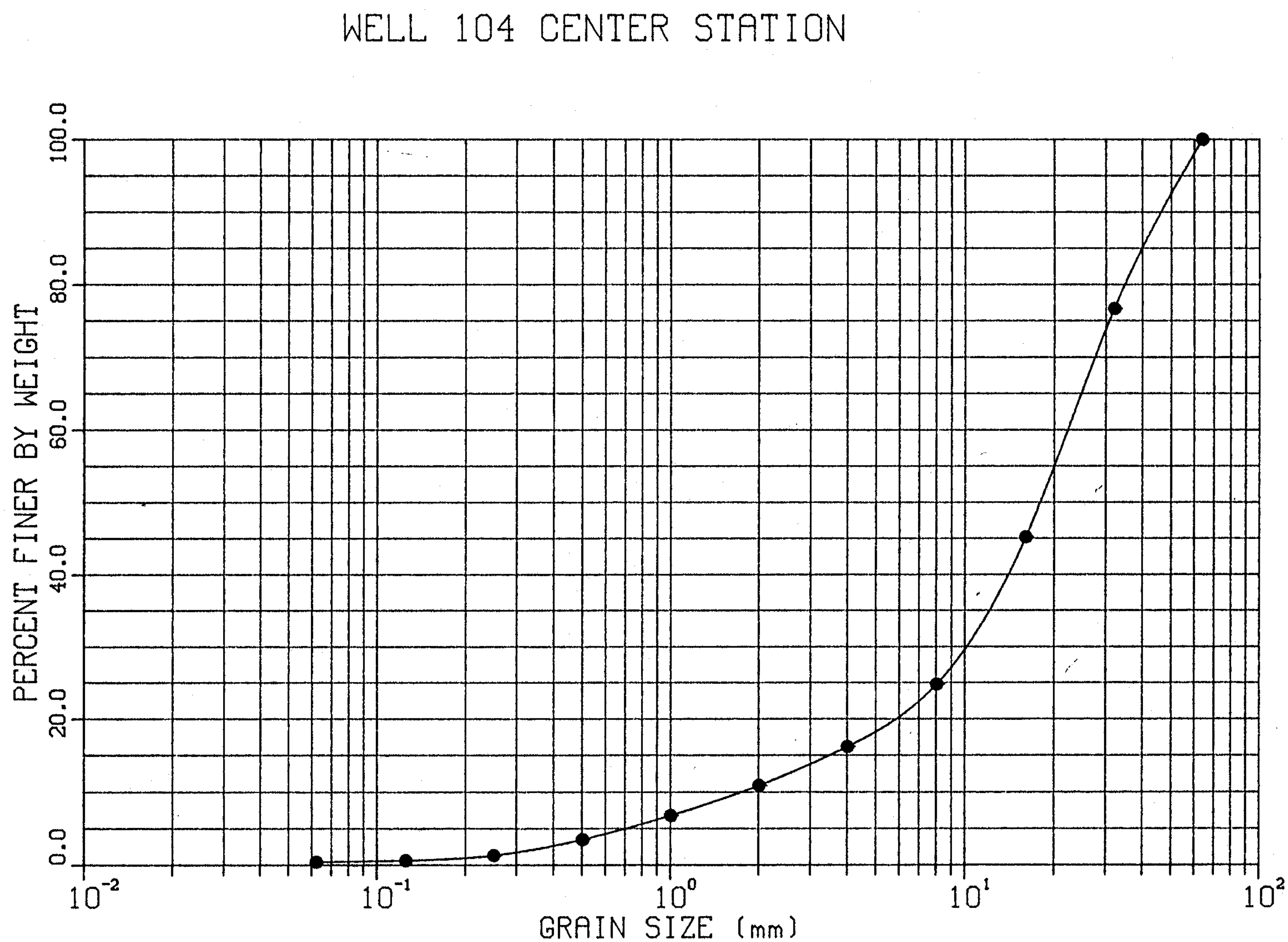
665 RUN
10-29-88
148 c.f.s.

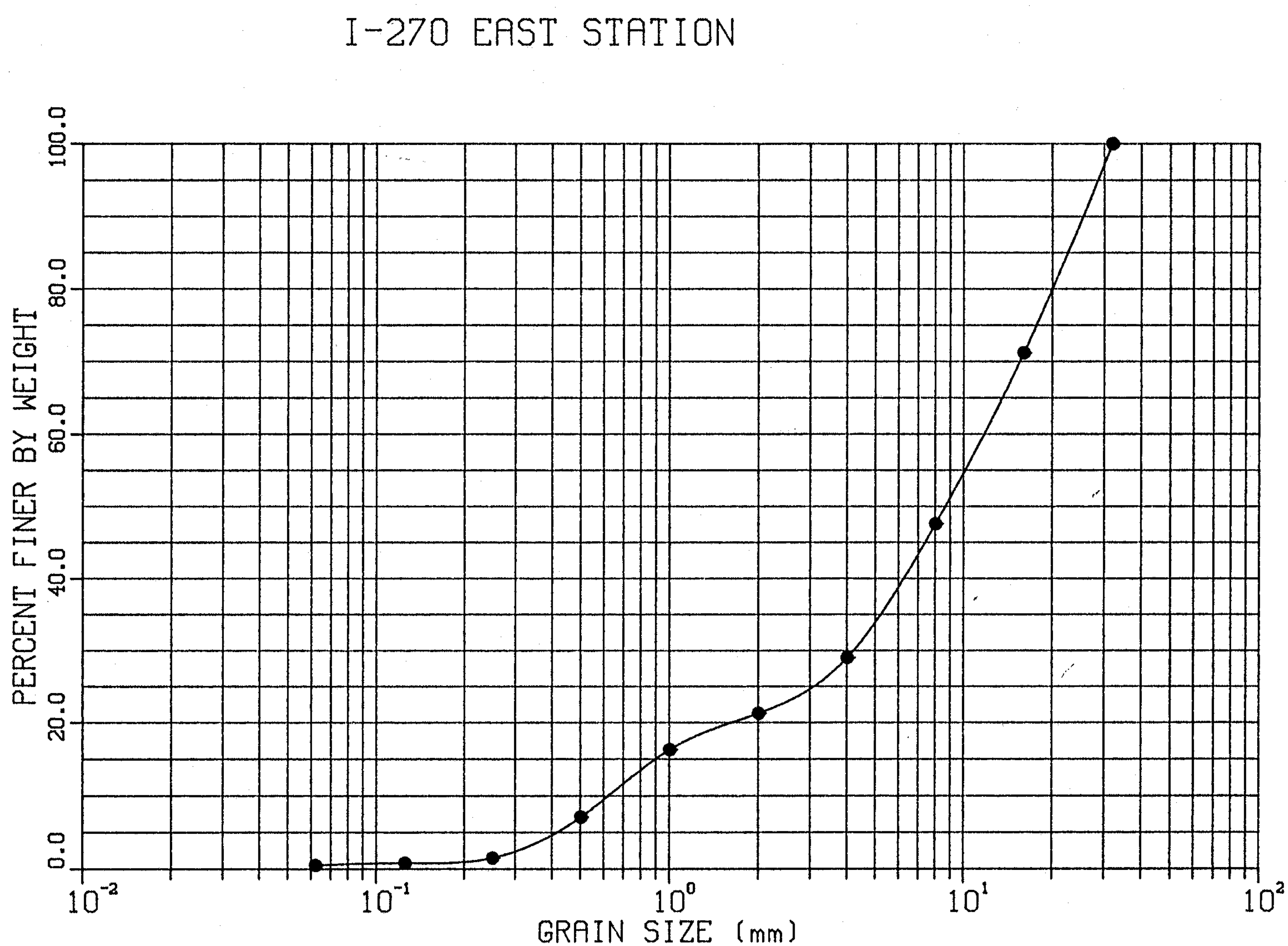
APPENDIX B

GRAIN SIZE DISTRIBUTION CURVES

WELL 104 EAST STATION

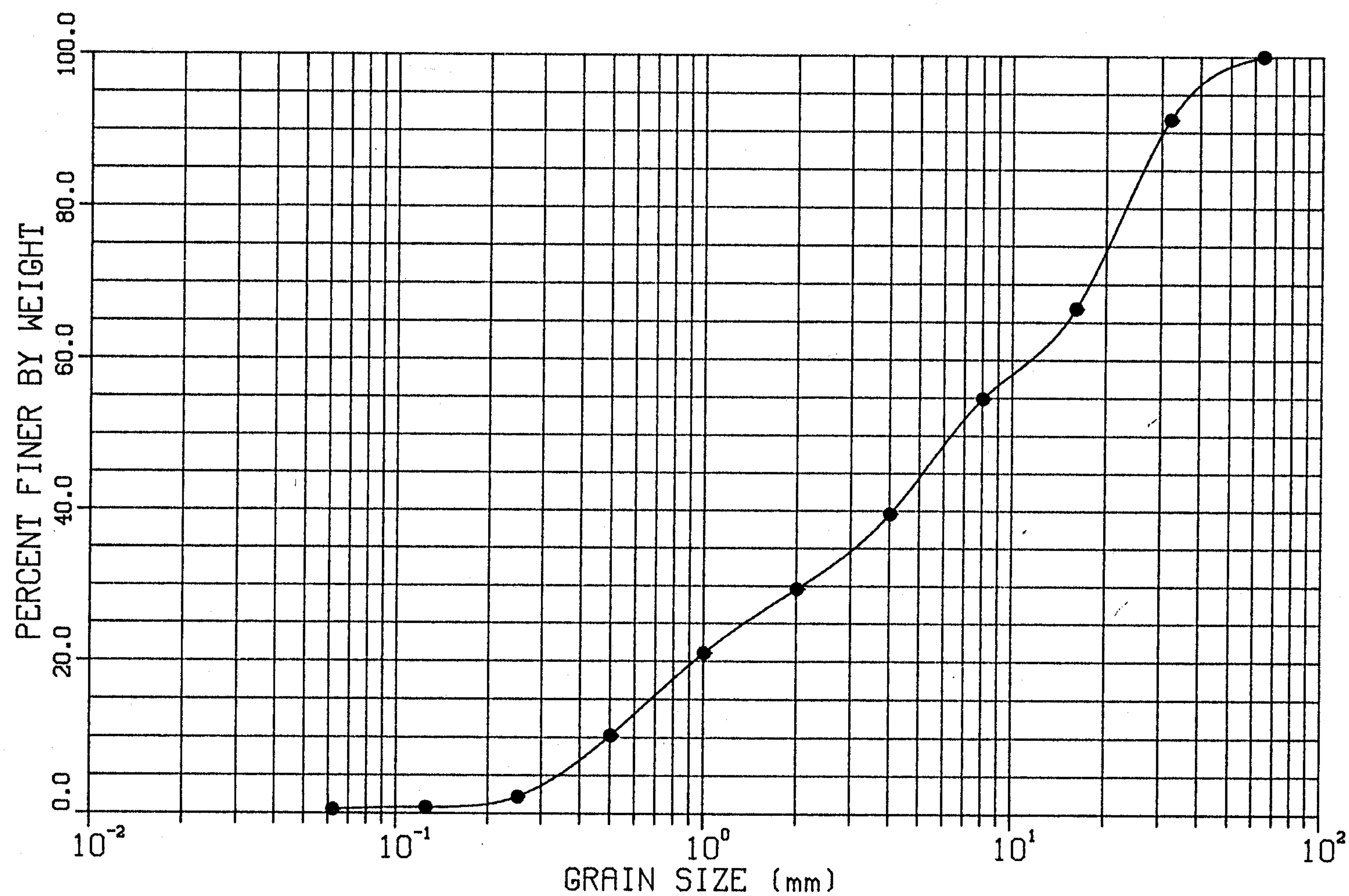



WELL 104 CENTER STATION
PERCENT FINER BY WEIGHT
100.0
80.0
60.0
40.0
20.0
0.0
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}
GRAIN SIZE (mm)

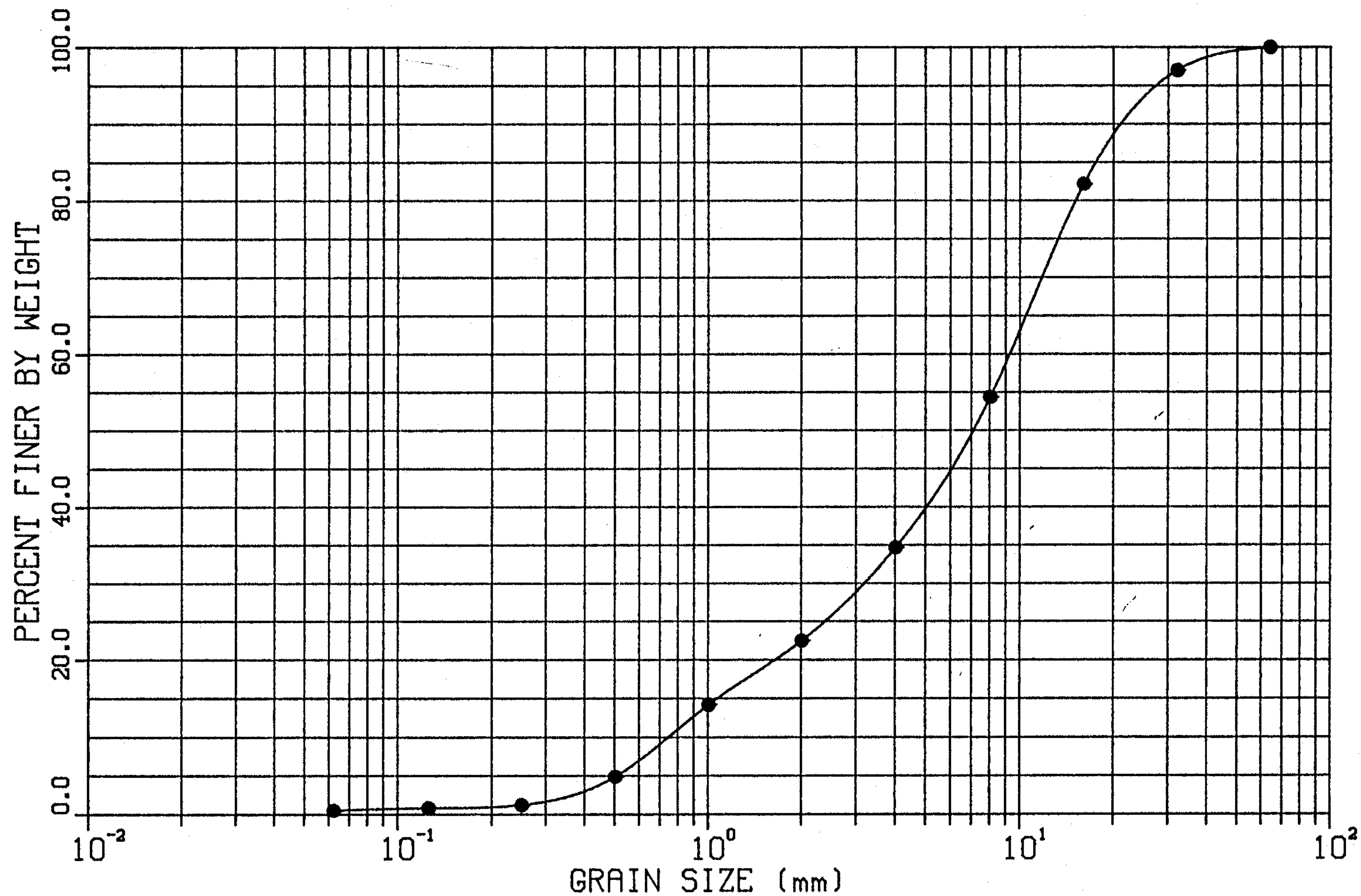


I-270 EAST STATION
PERCENT FINER BY WEIGHT
0.0
20.0
40.0
60.0
80.0
100.0
10^-2
10^-1
10^0
10^1
10^2
GRAIN SIZE (mm)

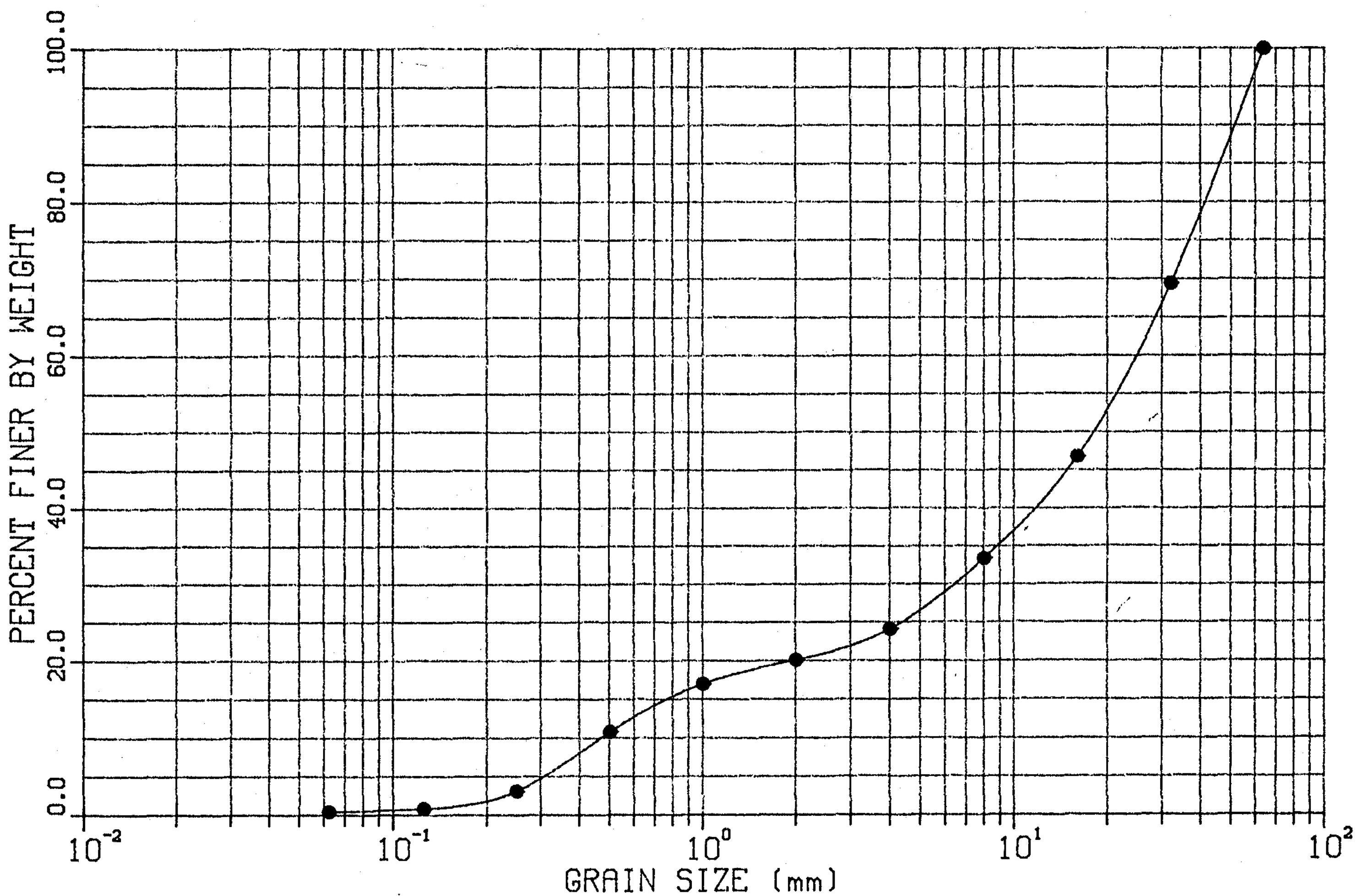
I-270 CENTER STATION



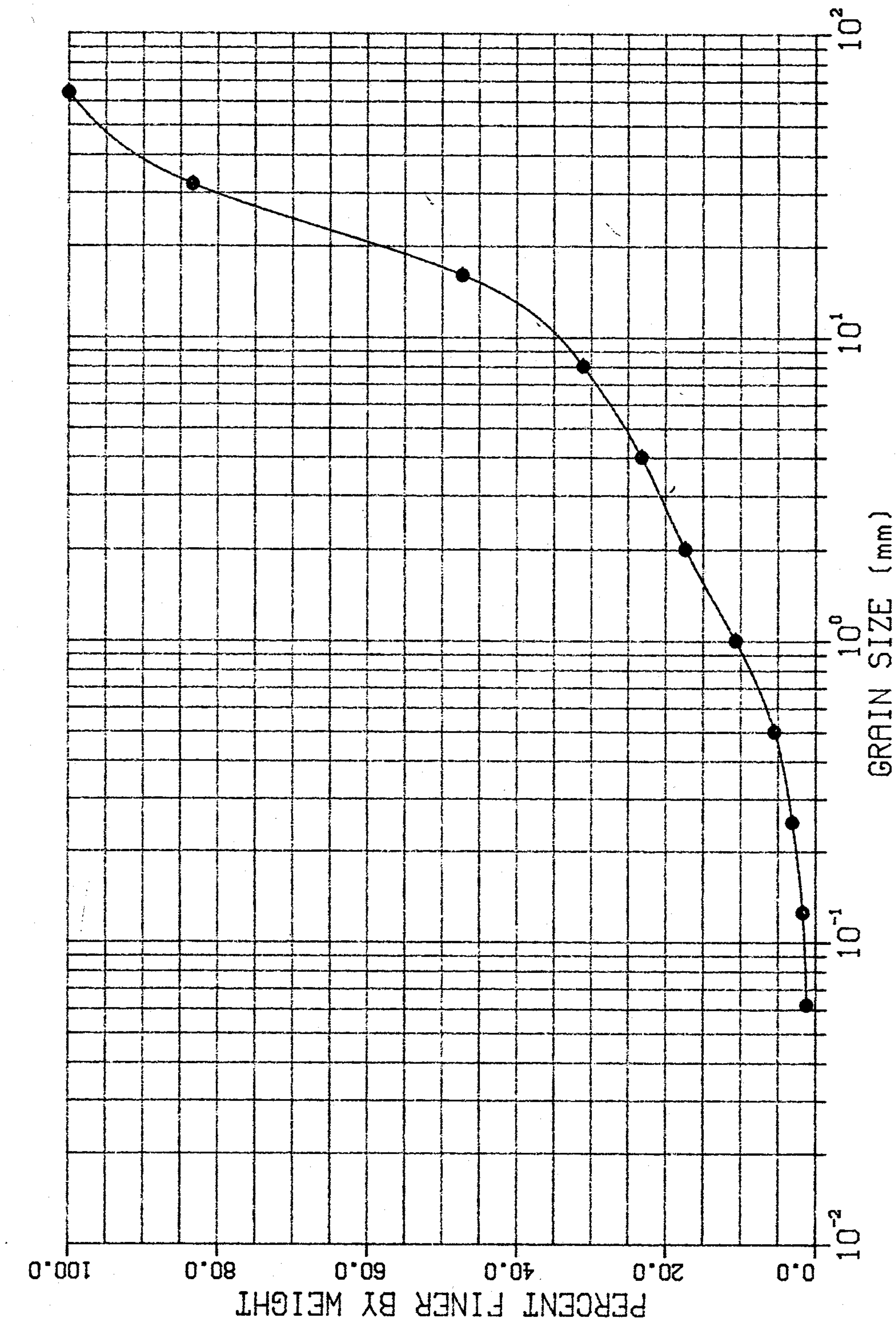
I-270 WEST STATION



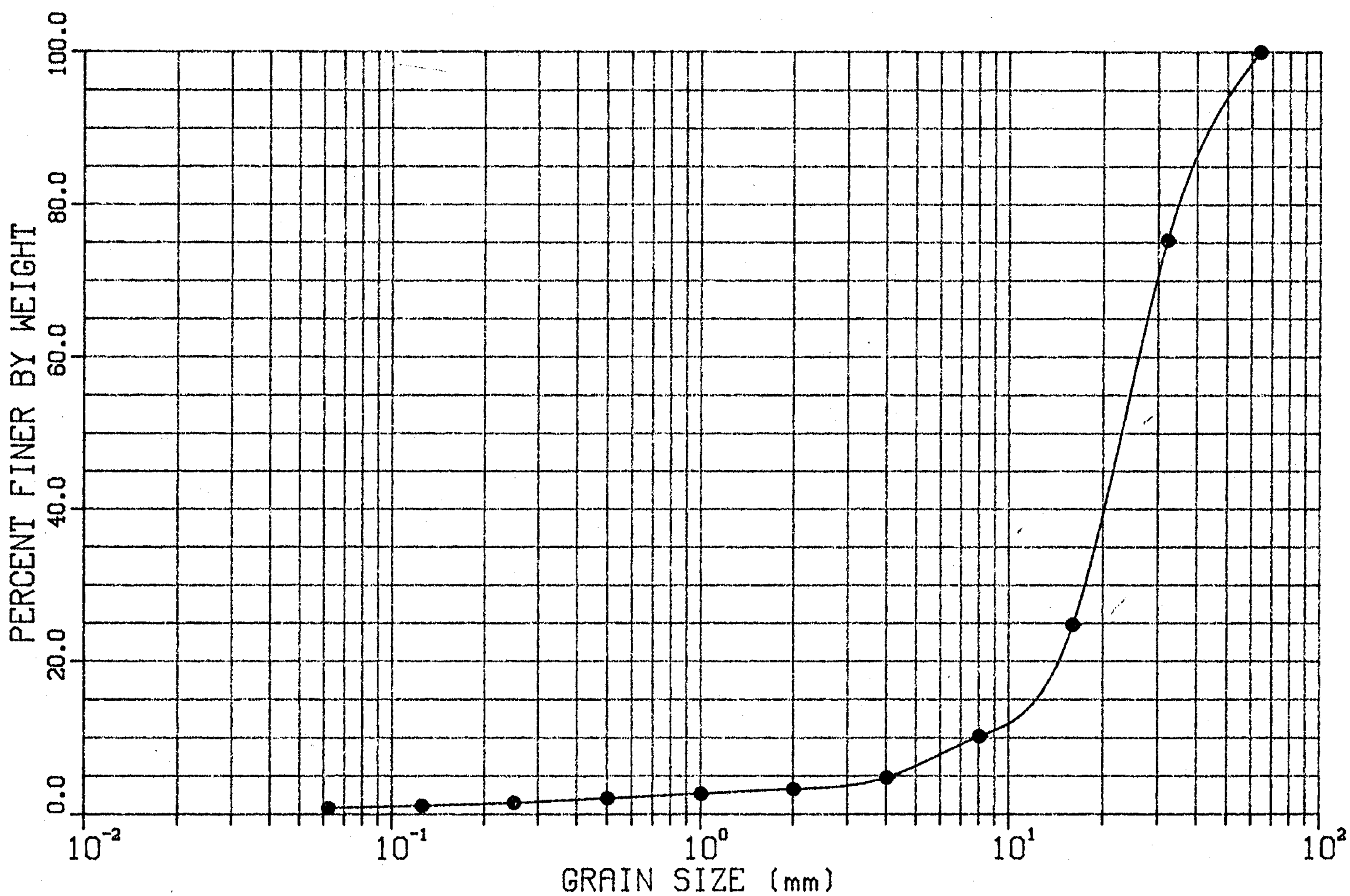
101 RUN: EAST STATION



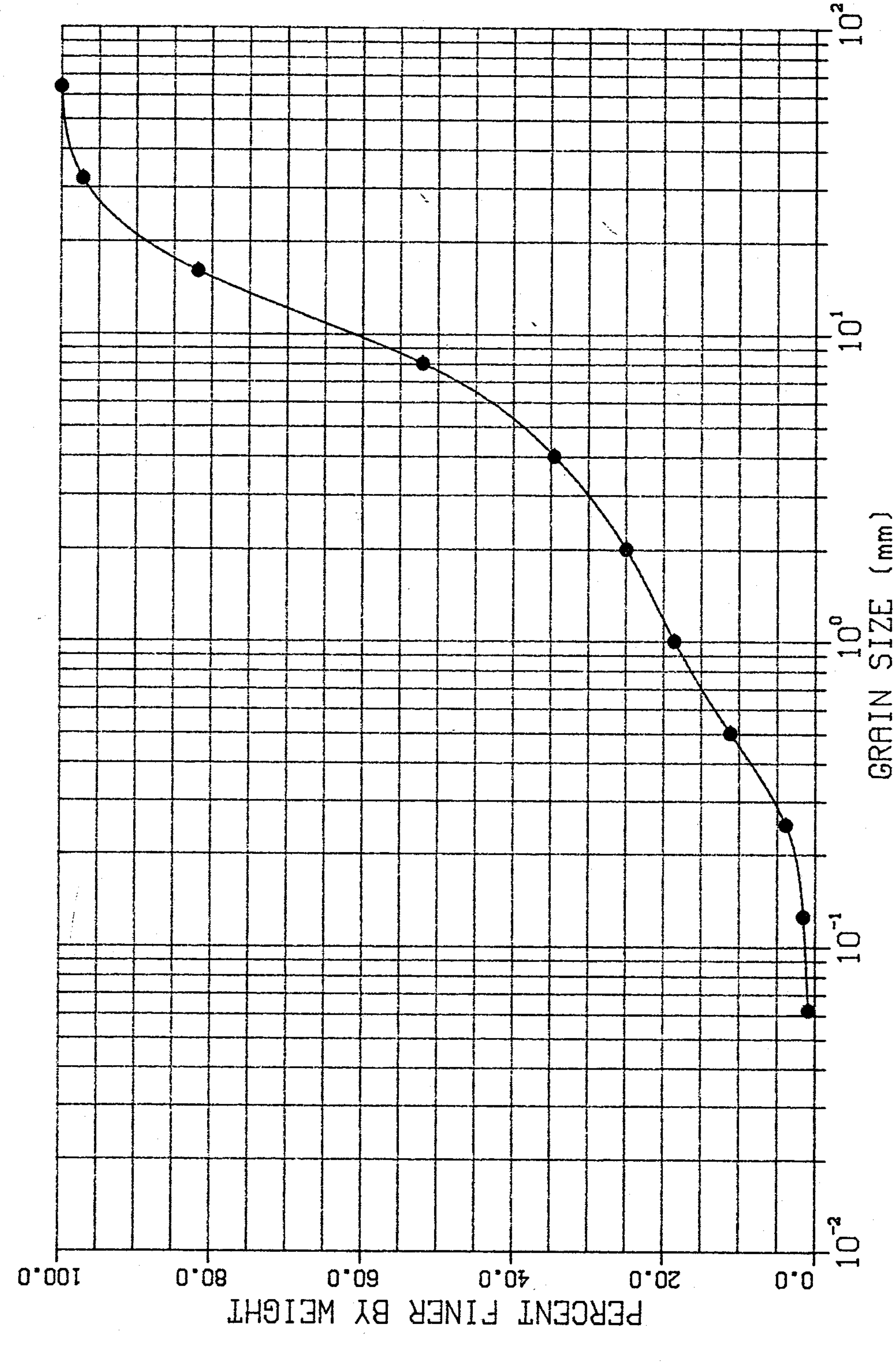
101 RUN: CENTER STATION



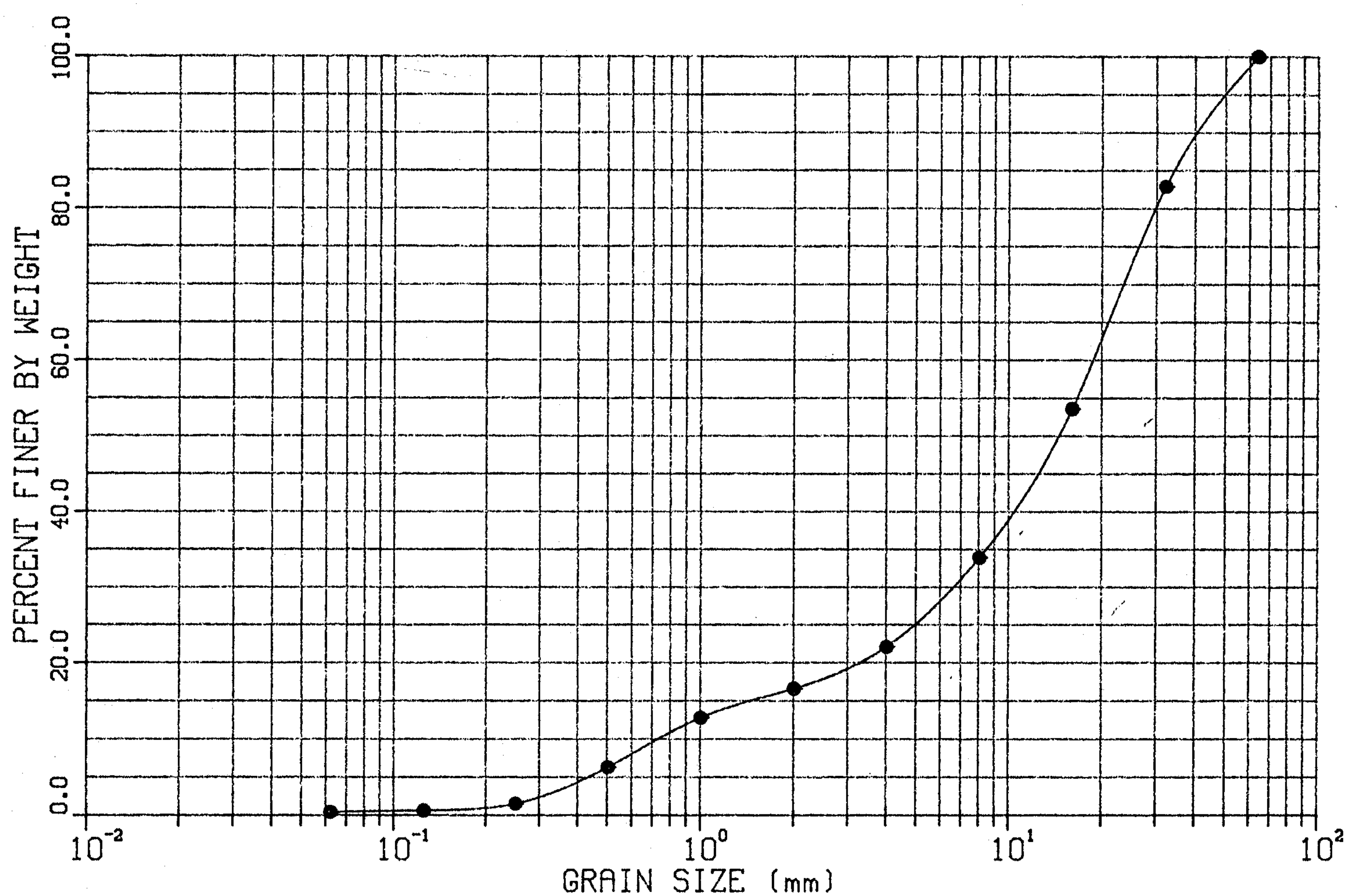
101 RUN: WEST STATION



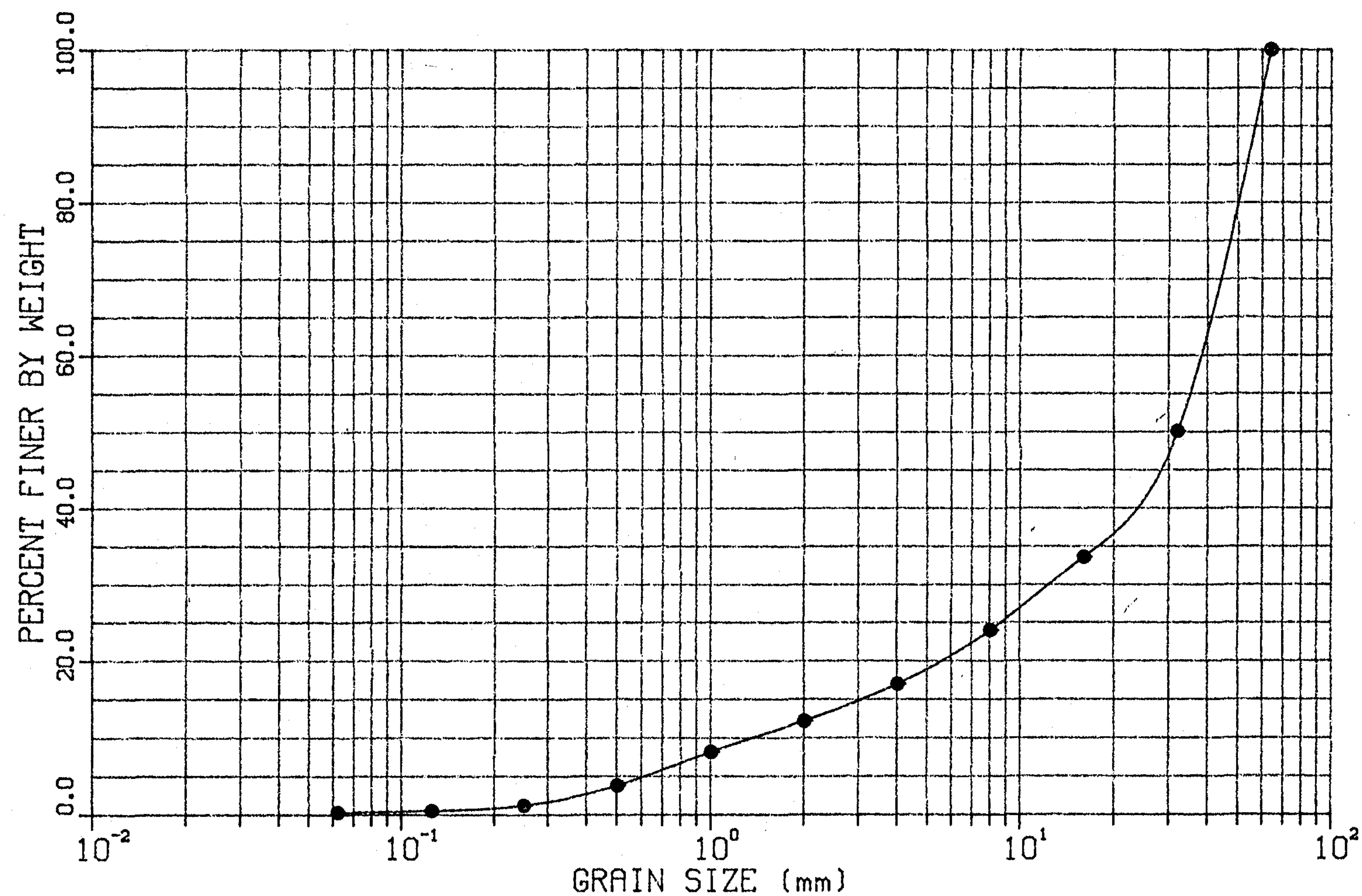
101 RIFFLE: EAST STATION



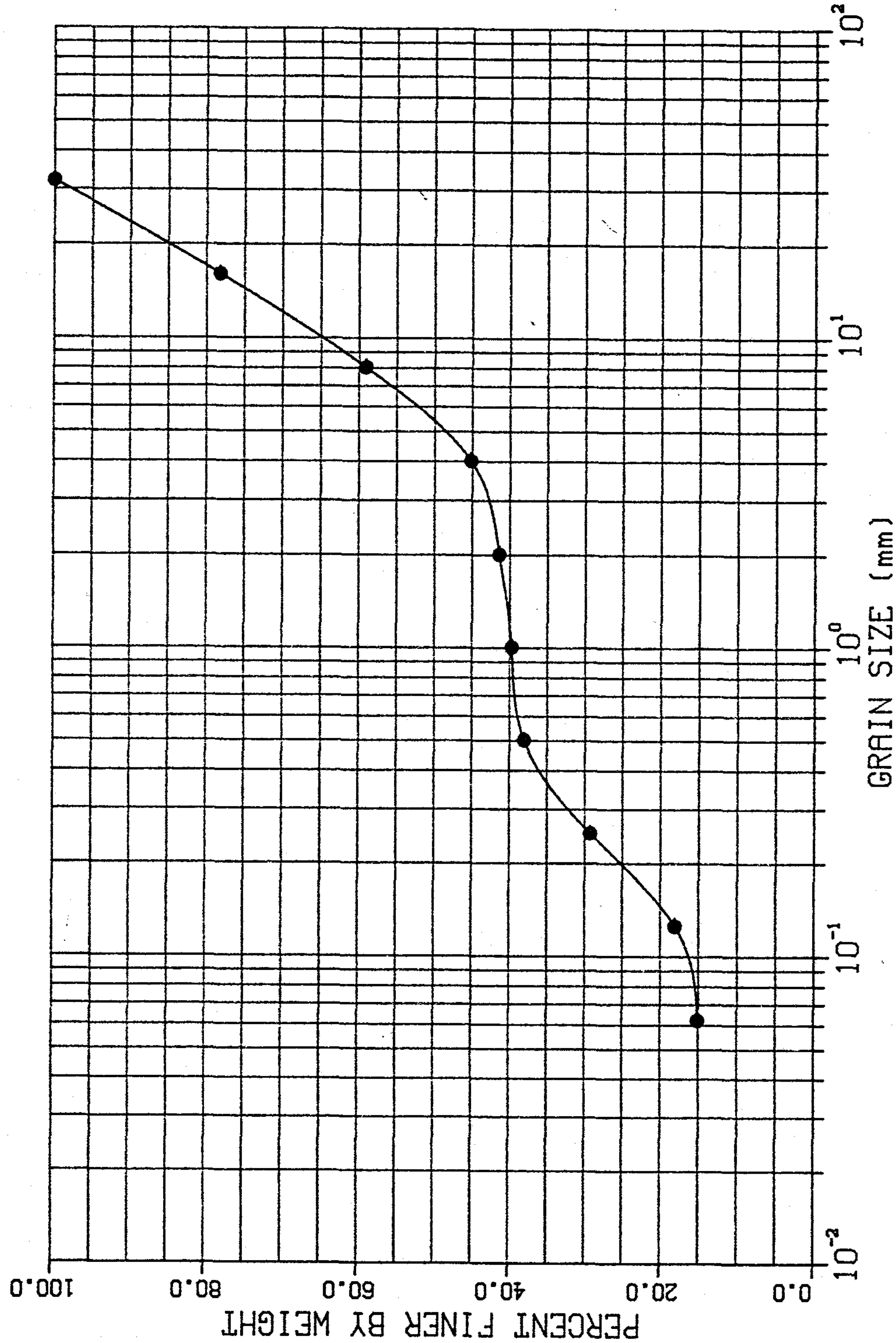
101 RIFFLE: CENTER STATION



101 RIFFLE: WEST STATION

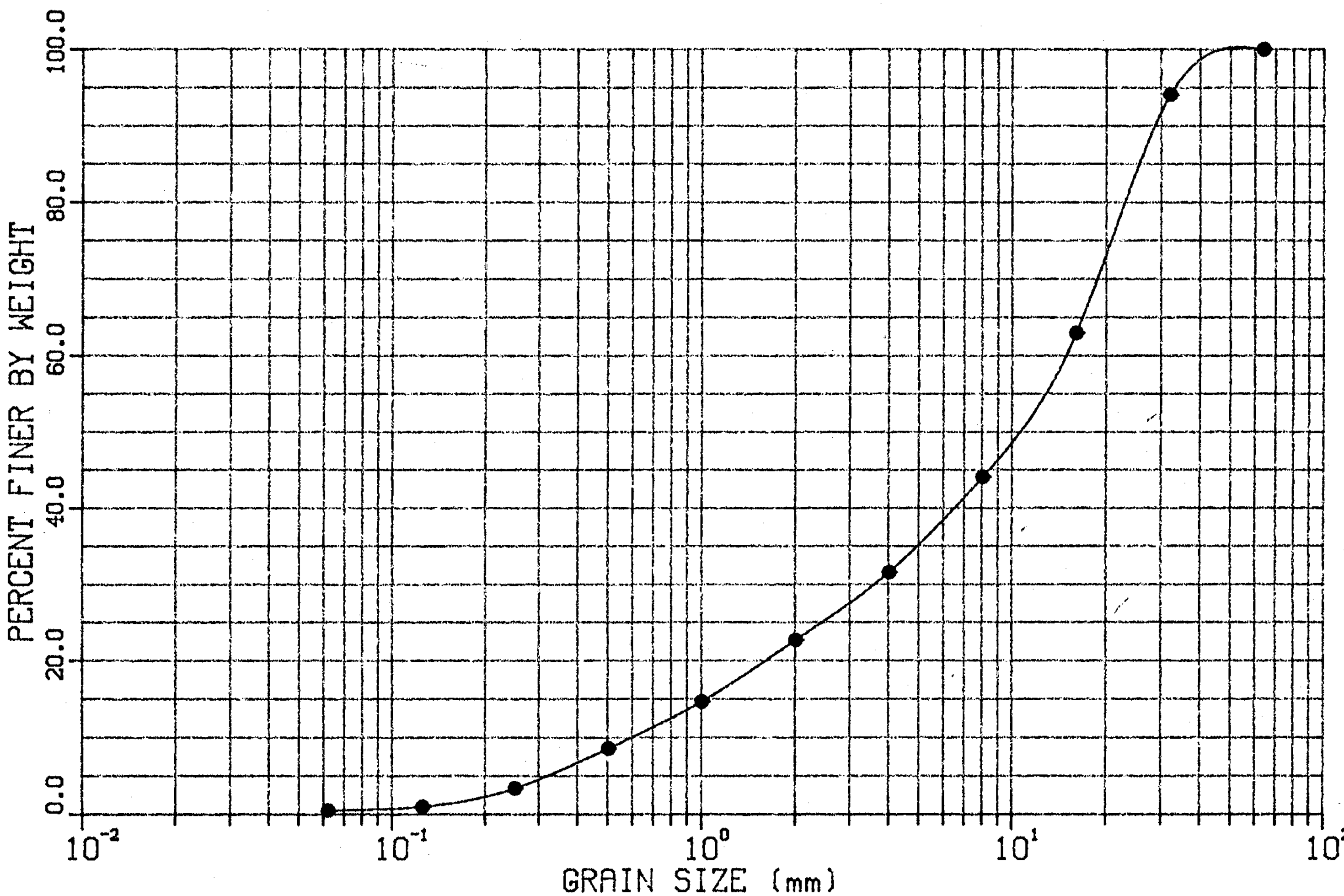


101 POOL

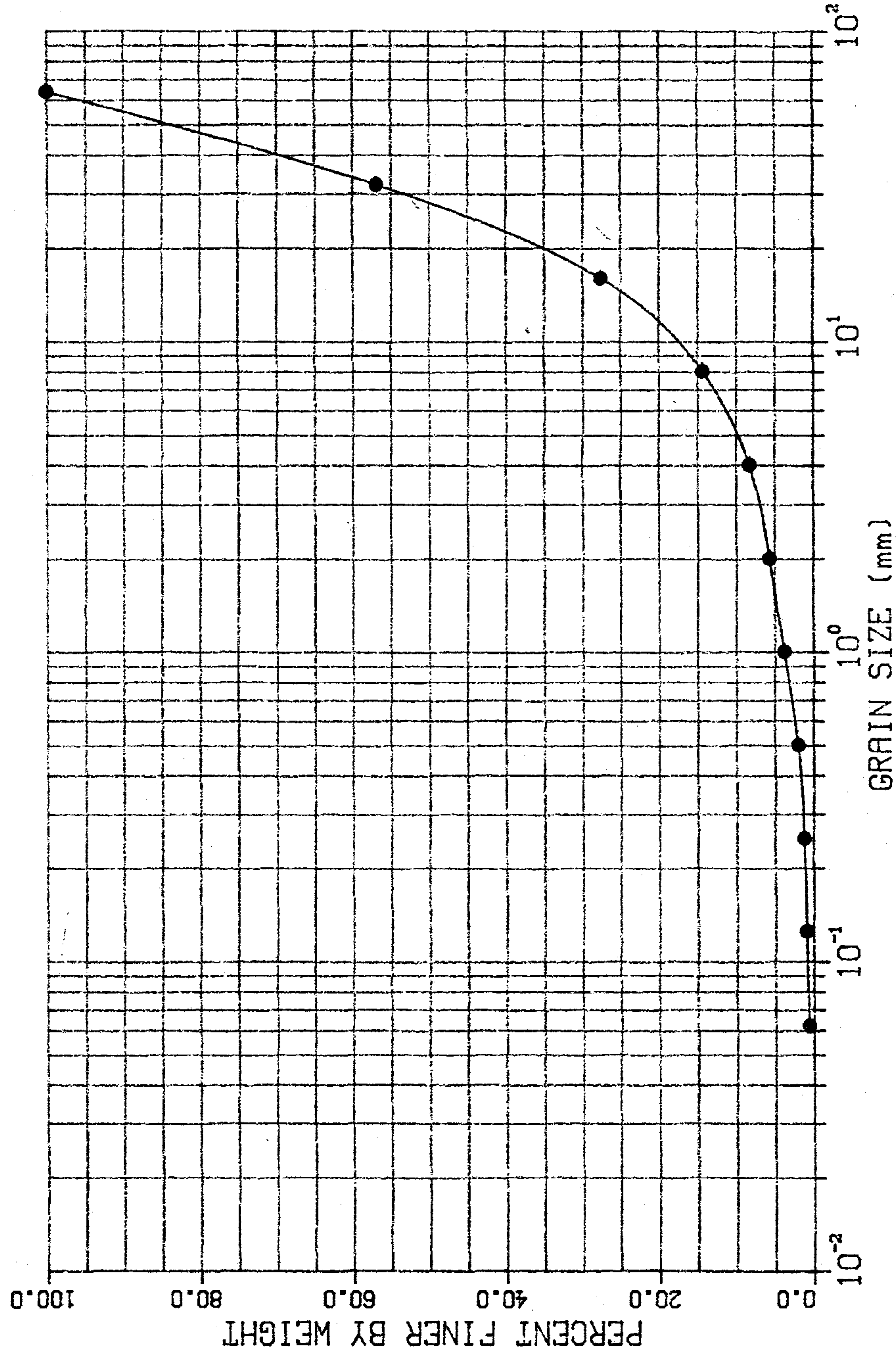


PERCENT FINER BY WEIGHT
100.0
80.0
60.0
40.0
20.0
0.0
GRAIN SIZE (mm)
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}

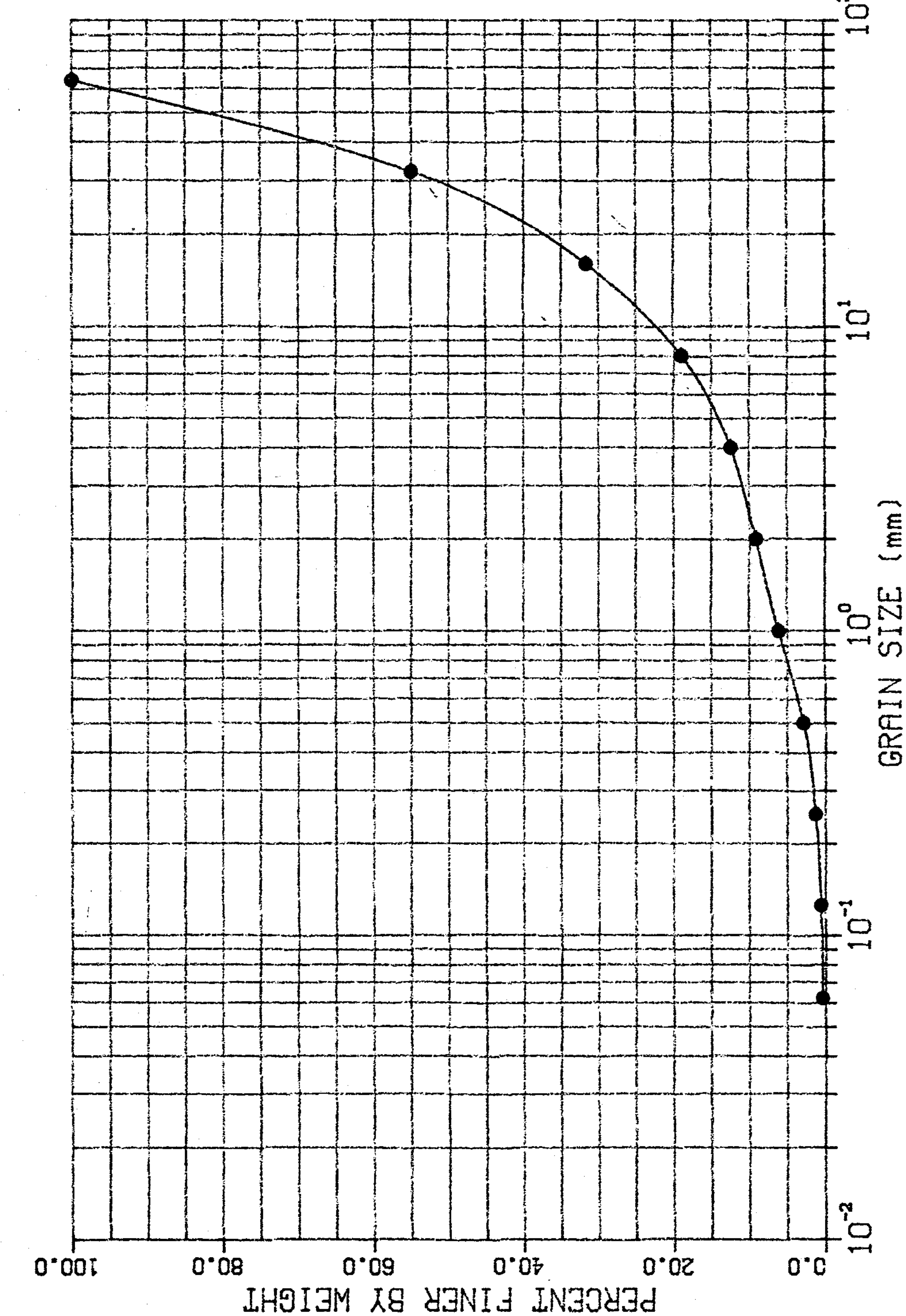
102 NORTH STATION



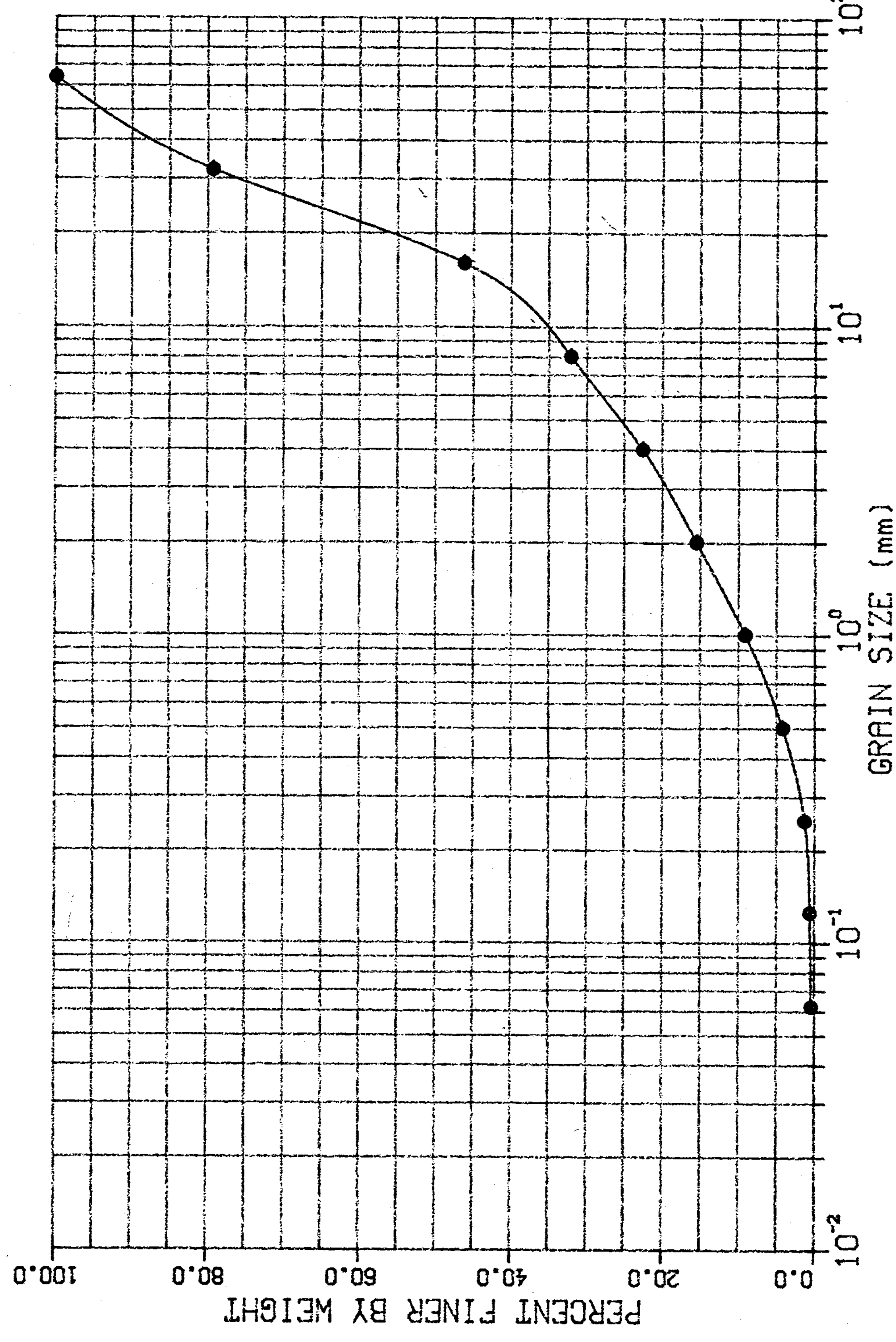
102 SOUTH STATION



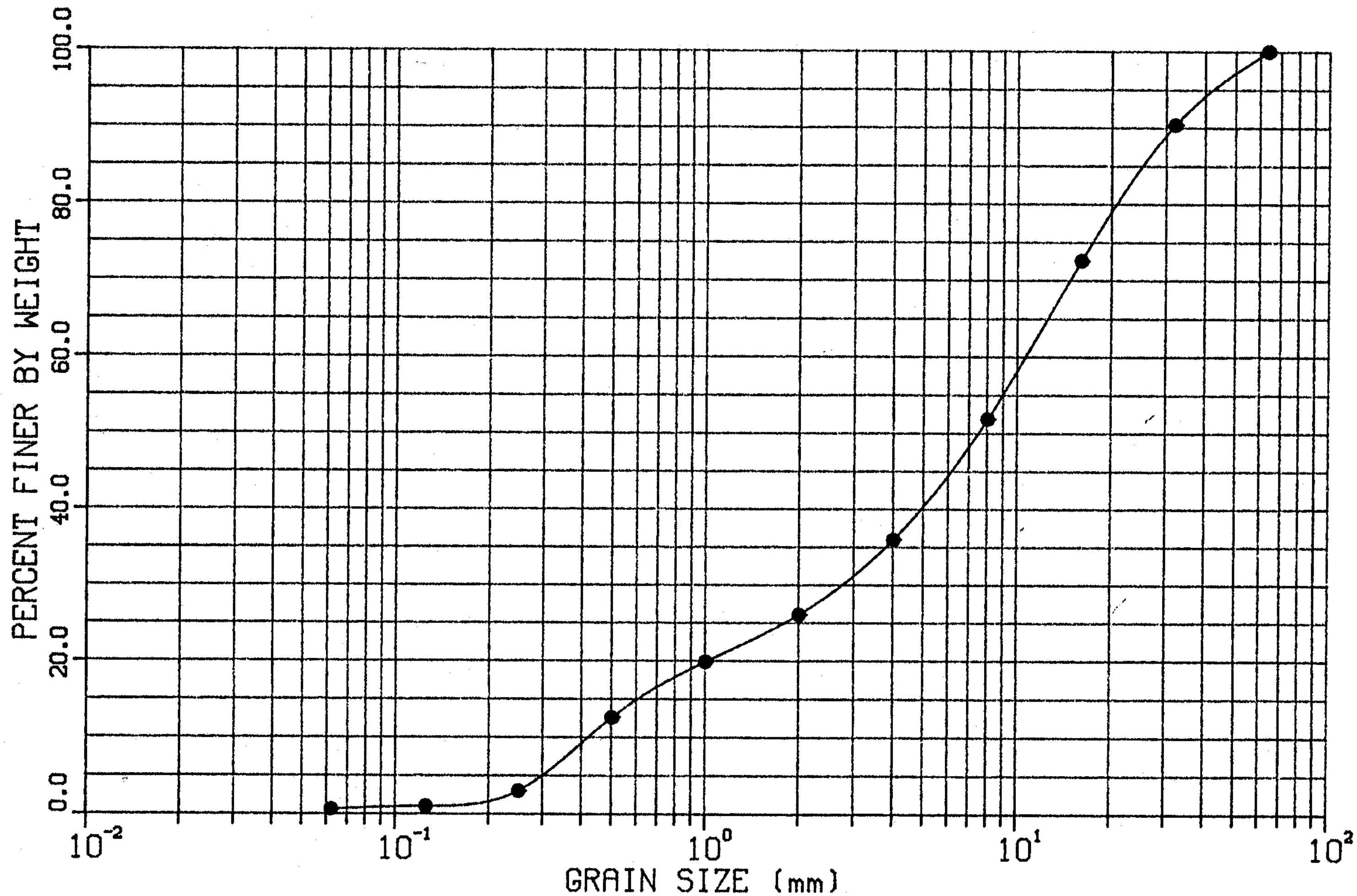
103 EAST STATION

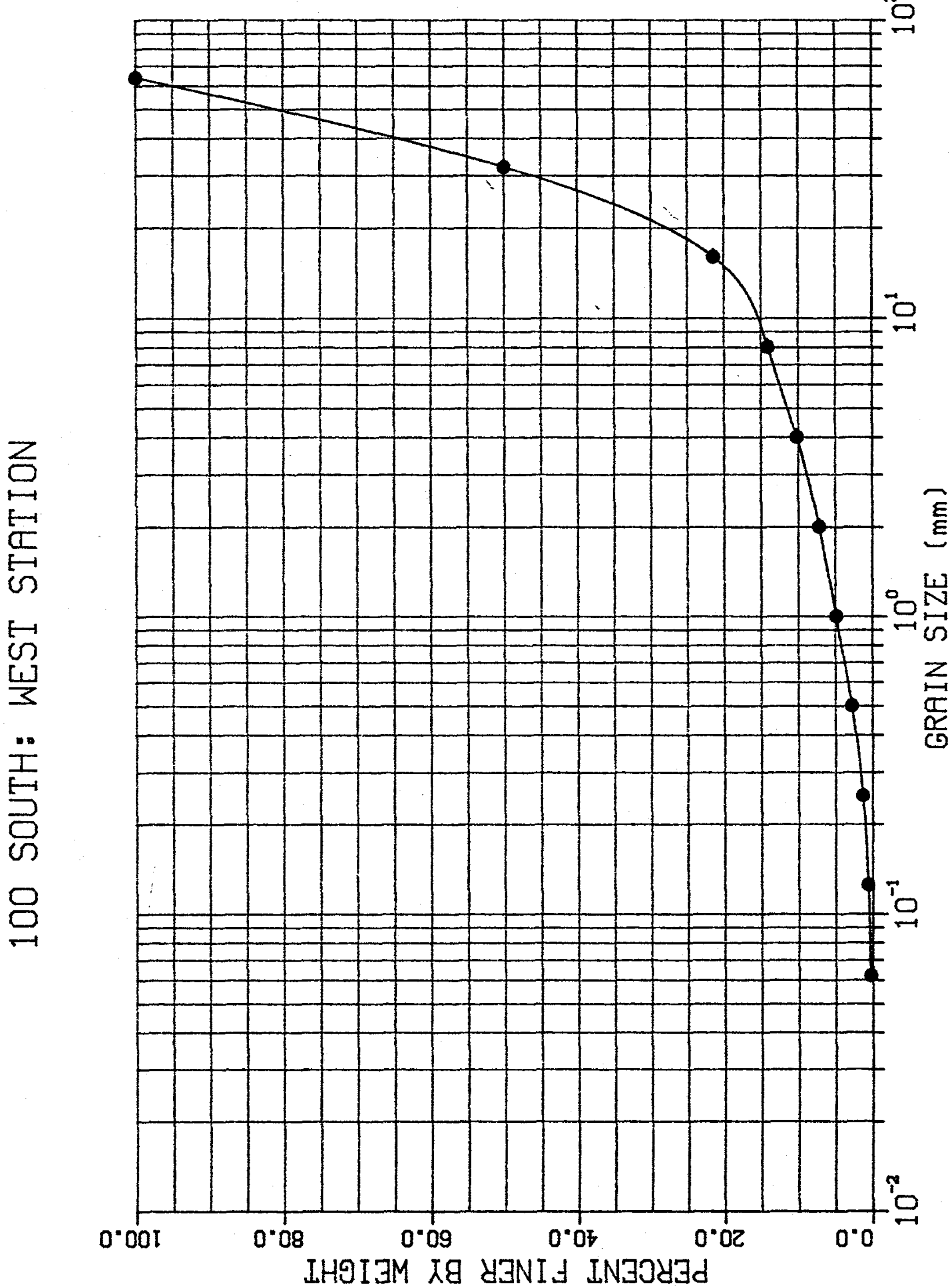


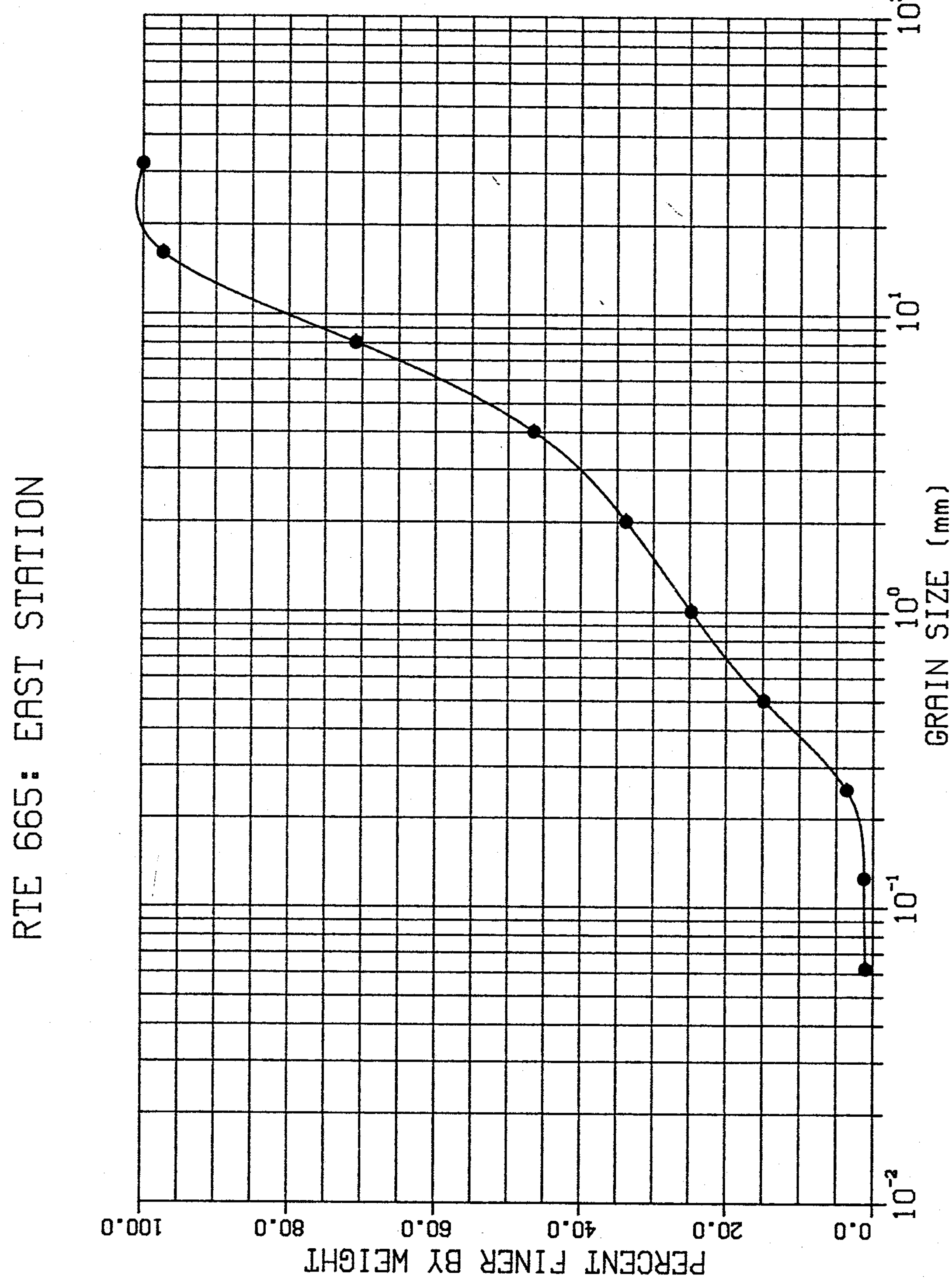
103 WEST STATION



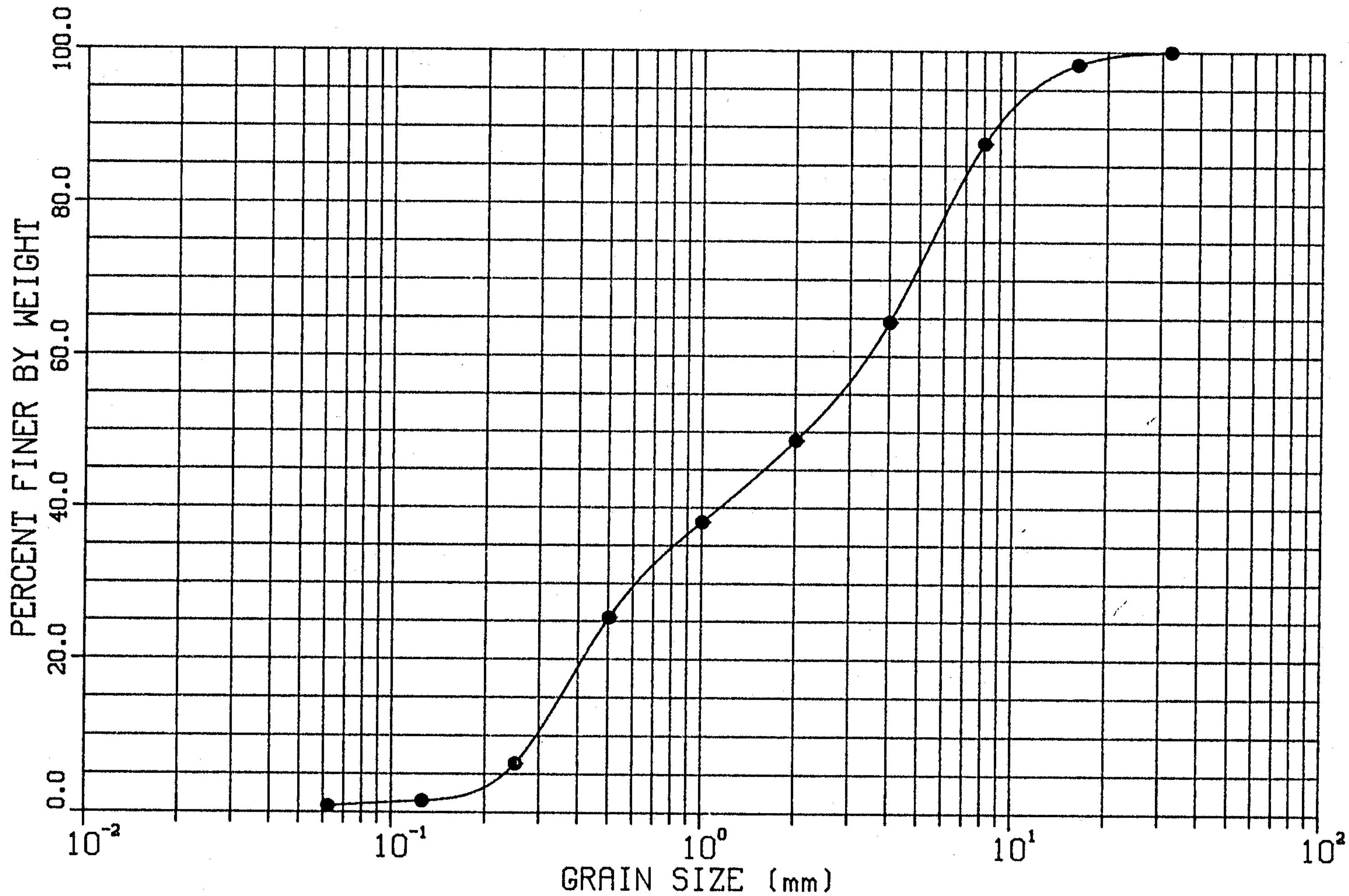
100 SOUTH: CENTER STATION

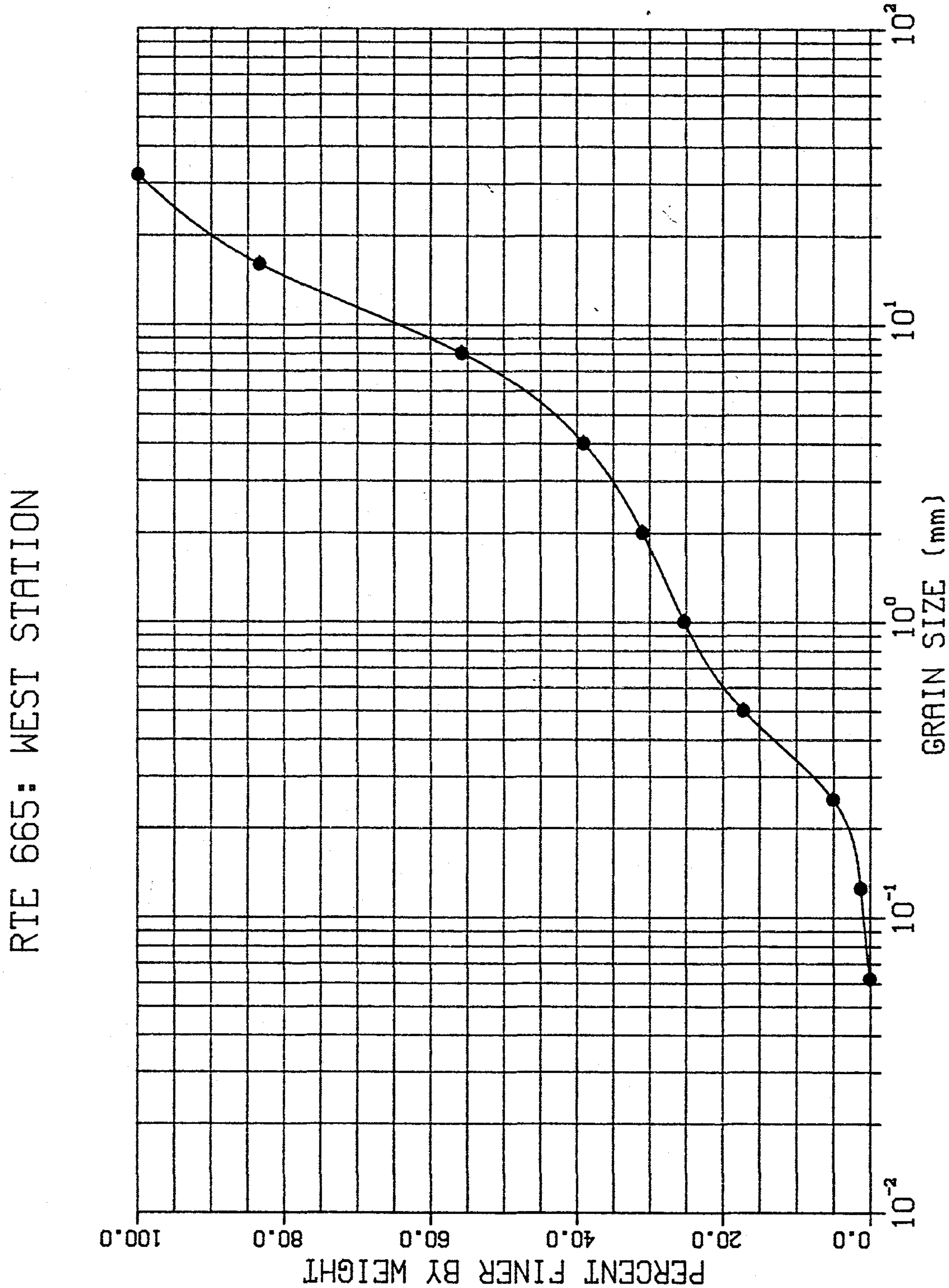



100 SOUTH: WEST STATION
PERCENT FINER BY WEIGHT
100.0
80.0
60.0
40.0
20.0
0.0
GRAIN SIZE (mm)
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}


RTE 665: EAST STATION
PERCENT FINER BY WEIGHT
GRAIN SIZE (mm)
0.0
20.0
40.0
60.0
80.0
100.0
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}

RTE 665: CENTER STATION




RTE 665: WEST STATION
PERCENT FINER BY WEIGHT
100.0
80.0
60.0
40.0
20.0
0.0
GRAIN SIZE (mm)
10^{-2}
10^{-1}
10^{0}
10^{1}
10^{2}

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